Abstract

Lake sediment archives are widely used in palaeoenvironmental reconstructions on the basis that lake ecosystems are sensitive to environmental changes and they integrate changes in the surrounding landscape and atmosphere (Birks & Birks 2006; Carpenter et al., 2007; Pham et al., 2008; Williamson et al., 2008). The primary aim of this research is to investigate the links between long-term changes in climate and changes in plant community dynamics, and the potential impacts of these interactions on catchment palaeo-erosion rates during the Holocene in a high relief lake catchment. There is a distinct lack of palaeo-record coverage for the Hengduan Mountains and surrounding areas including the south eastern edge of the Tibetan Plateau (Zhang & Mischke, 2009). The Lake Mugecuo MG1 record analysed here is one of the few high resolution Holocene palaeoenvironmental and palaeoecological records available for Southwestern China and one of the first for this part of the Hengduan Mountains in Sichuan Province, extending back to ~12.6 kcal.yr BP.

The long-term relationship between major climatic change, vegetation change, and denudation events is complex and poorly understood (Willis et al., 1997). A PFT (plant functional type) approach was adopted to provide a method of classifying pollen taxa into plant communities in order to determine whether plant functional diversity mediates the effects of climate on lake catchment dynamics (including long-term erosion rates) at Lake Mugecuo. $^{10}$Be concentrations in bedrock and fluvial sediment samples were used as an independent proxy to estimate surface erosion rates, along with lacustrine deposits of quartz from two shallow water cores (MG3 & MG4) which are used to estimate palaeo-erosion rates for the Lake Mugecuo catchment. A published speleothem record (Wang et al., 2005; Dykoski et al., 2005) from Dongge cave (25°17′N, 108°5′E; 680 m), is used as an indicator of the general pattern of effective moisture change in this region during the Holocene in order to determine if there is a direct link between climate change and the lake deposition record and whether factoring in vegetation improves our understanding of the link between climate change and the lake deposition archive.

The MG1 record provides a unique insight into the environmental and vegetation changes that took place during the Late-Glacial/Early Holocene transition. Statistical analysis of the data reveals that plant functional diversity and lake catchment dynamics (using particle-size analysis as an insight into catchment stability) respond to long-term changes (>2,800yr) in climate on in the Lake Mugecuo catchment whilst interactions between plant functional diversity and lake catchment dynamics are found to vary on shorter timescales (<2,800yr). It is argued here that apparent environmental shifts captured in the Lake Mugecuo record are primarily driven/mediated by changes in plant functional diversity and that an array of different types of species and/or functional groups play an important role in modulating and landscape evolution across millennial timescales (approximately 1,000-3,000 years). Vegetation is found to have a filtering effect on the signals of the proxies that are traditionally used as indicators for changes in erosion and deposition in palaeoenvironmental reconstructions (e.g. particle size). $^{10}$Be cosmogenic measurements provided supplementary data on past changes in landscape dynamics in the Mugecuo catchment which the conventional proxies used in palaeoenvironmental studies were unable to achieve.
I would like to thank the following people for their help with this research:

- My supervisors Dr Liam Reinhardt, Dr Richard Jones and Dr Mike Ellis for their contributions to this research
- Dr Enlou Zhang and colleagues at the Nanjing Institute of Geography and Limnology for logistical and technical support whilst on fieldwork in China
- Laboratory staff in the Geography Department at Exeter University for their assistance and guidance
- The Quaternary Research Association (QRA), the Royal Geographical Society (RGS) and the NERC NRCF-Steering Committee for providing additional funding for laboratory and fieldwork costs

A special thanks must also go to my family and friends for their unconditional love, support and enthusiasm throughout this process.
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Table 5.10 Integration times for each of the palaeo-, modern fluvial and bedrock erosion rates for the Lake Mugecuo catchment.

Table 5.11 Summary of the environmental, vegetation, and climatic characteristics of each multi-proxy zone inferred from the Lake Mugecuo proxies.
1.1 Research Context

Landscapes are shaped by the interplay of hydrologic, biologic and tectonic processes operating over a wide range of space-time scales. The mountain regions of China provide rich habitats for much of the country’s remaining biodiversity, owing to the heterogeneity of climates and soils, rapid elevational changes, varying aspects of slope direction, abundant microhabitats, and limited suitability for cultivation (Körner and Spehn, 2002). The purpose of this research is to reconstruct past vegetation dynamics and past variation in sediment dynamics at a remote high alpine lake in Southwest China in order to determine whether vegetation mediates the effects of climatic change on lake catchment dynamics. Multi-proxy studies are becoming increasingly popular in palaeo-ecology, palaeo-climatolology and palaeolimnology due to the potentially independent lines of evidence they offer for environmental reconstruction (Lotter, 2003). A sedimentary archive is used here to determine the effects of plant functional diversity on catchment dynamics through time in order to assess whether plant functional diversity mediates the effects of climate on lake catchment dynamics (including long-term erosion rates). Plant functional types (PFTs) are defined as groups of plants exhibiting either similar responses to an environment or similar effects on major ecosystem processes (Noble and Gitay 1996; Skarpe 1996; Diaz and Cabido 1997; Duckworth et al. 2000; Kleyer 2002; Lavorel and Garnier 2002). The concept of plant functional types can be applied at a range of spatial scales, from the plant community, through the ecosystem and landscape, to global scales, with the types of attributes that are important dependent on the scale in question (Duckworth, 2000). In addition, the use of PFTs has provided an ecological basis for the treatment of plants from different regions in a compatible way whilst reducing the number of entities involved (Prentice and Webb, 1998).

Numerous studies have shown that long-term erosion rates do not correlate with mean annual precipitation and temperature, implying that long-term erosion rates are not sensitive to catchment hydrology (von Blanckenburg, 2005). This observation suggests that a change in precipitation patterns does not per se change catchment dynamics. However, given that changes in precipitation and temperature can alter the ecology of any region (Stenseth et al., 2003), it is intuitively reasonable to suggest that climate change indirectly perturbs a landscape through its influence upon a landscape’s ecosystem (Reinhardt et al., 2010). Figure 1.1 represents the conceptual model that forms the basis of this research by outlining how biological activity likely mediates the effects of climate on soil ecosystem functioning i.e. the erosion/deposition of sediment through a system. The basis of this conceptual model is further discussed in the following section (1.1.1).

The quantification of spatial and temporal patterns of erosion in high-relief terrain has been made more tractable through the use of cosmogenic nuclide techniques which permit the direct measurement of long-term erosion rates in both in-situ, and sediment samples (Bierman and Steig, 1996). The reconstruction of past variation in sediment dynamics at a dynamic high alpine lake in this study is based on cosmogenic nuclide measurements along with an array of organic and lithological proxies in order to determine the effects of plant functional diversity on catchment dynamics through time, and to determine whether vegetation mediates the effects of climate on landscape dynamics through time. The relationship between plant functional diversity and geomorphology is a diverse and rapidly developing field which is reflected in
the scope and nature of this research. The following sections provide a literature review of the central concepts behind this study.

Figure 1.1 Conceptual diagram. Atmospheric and climate change can impact soil ecosystem functioning by directly affecting biological activity and composition of soil communities, (pathway 1) or by affecting individual plant functioning (e.g., root biomass production) (pathway 2). Atmospheric and climate change can also impact soil ecosystem functioning through shifts in plant community composition (pathway 3), which can result from changes in individual plant function (altered plant-competitive interactions) or from changes in individual plant distribution (species range shifts) which will thus affect the erosion/movement of sediment through a system. Adapted from Kardol et al., 2010.

1.1.1 Geomorphology and lake catchment dynamics

Geomorphology, the study of landscape change, stands in the centre of an emerging science of the Earth's surface, where strong couplings link human dynamics, biology, biochemistry, geochemistry, geology, hydrology, geomorphology and climate change (Murray et al., 2009). Geomorphology seeks to explain the diversity, origins and dynamics of landscape change based on a) the quantitative characterization of terrain, and b) an ever-improving theory that describes the modification of topography by the many and varied processes that sculpt it (Tucker & Hancock, 2010). Climate change affects physical components (size and frequency of floods, sediment fluxes) and biological components (sediment-stabilizing vegetation, destabilizing macroscopic animals) of the Earth's surface systems, which can transform landscape and ecosystem regimes and therefore, also the feedbacks between them (Murray et al., 2009; Reinhardt et al., 2010). Such changes represent perturbations to landcapes and their biological communities and may lead to large-scale changes in patterns of erosion and deposition. Whilst it is widely accepted that physical processes and landforms drive the evolution of organisms and ecological successions (e.g., Pickett and White, 1985; Steiger et al., 2005), it is also becoming recognised that living organisms can significantly modify geomorphic processes and thus have a strong control on landform dynamics (e.g., Howard and Mitchell, 1985; Naiman et al., 1988; Butler, 1995; Corenblit and Steiger, 2009; Reinhardt et al., 2010)). However, there is limited understanding of the mechanisms by which biogeomorphic processes contribute to defining earth surface landscape dynamics through reciprocal adjustments between landform dynamics, population dynamics, ecological successions and organism evolution.

The long term relationship between major climatic change, vegetation change, and denudation events is complex and poorly understood (Willis et al., 1997). Geomorphologists studying hillslopes and drainage basins have long recognized that biology affects sediment production, transport and landscape morphology
The pioneering works of Clements (1928), Cain (1944) and Braun Blanquet (1932) had the view that climate was the primary variable affecting plant distributions, with edaphic and topographic variables exerting a secondary influence (Marston, 2010). The pioneering works of geomorphologists such as De La Noë and De Margerie (1888), Tricart and Cailleux (1969, 1974) all recognised that climate acts on geomorphic processes by acting on soil, vegetation, weathering and runoff. Climatic change is typically the primary variable/driving force behind change (Marston, 2010) but the question remains; how is this related to vegetation change and changes in the physical structures of a landscape? The relationship between climatic cause and depositional-effect has been challenged in various modelling studies (Coulthard and Van de Wiel, 2007; Jerolmack and Paola, 2010; Van de Wiel and Coulthard, 2010; Ganti et al., 2011). These studies argue that sediment transport through catchments is controlled by internally generated critically self-organised (SOC) processes that act as non-linear filters that ‘shred’ externally generated signals such as climate change and that part of this filtering process is the ‘smearing’ effect of stochastic sediment transport (Jones et al., 2013). The term ‘fluvial filtering’ has been proposed to refer to the role of the cascade of biotic and sediment storages located along river systems in trapping nutrients that might otherwise reach a lake bed (Meybeck and Vorosmarty 2005). Such filtering can result in a ‘smoothing’ of the signal of an individual proxy in a palaeoenvironmental reconstruction and fluvial filtering processes have the power to greatly influence lake sediment that is archived in a proxy sequence.

Unfortunately, our poor understanding of the non-linear relationships between precipitation/temperature and erosion/sediment-deposition means that we cannot yet predict the impact of climatic change on these landscape processes. Long-term erosion rate studies (>10^3 yr) have shown that there is no direct relationship between mean annual temperature or precipitation and catchment erosion (von Blanckenburg, 2005), while shorter term (<10^2 yrs) field studies evidence a host of often contradictory relationships (Langbein and Schumm, 1958; Walling and Webb, 1983). In the short term, changes in climate and vegetation influence sediment delivery rates, particularly in semi-arid regions (e.g. Knox 1983). Climate change can alter sediment storage by triggering mobilization or deposition on flood plains, colluvial hollows (Reneau et al. 1990), and terraces (Bull, 1964), thus substantially altering short-term rates of sediment delivery. However, measurements of short-term delivery (e.g. Riebe 2001), do not clearly reveal how climate affects erosion rates because sediment delivery rates reflect both the rate of sediment production by erosion and any changes in sediment storage (Trimble 1977). Importantly however, these short term studies have highlighted the crucial role played by vegetation cover on landscape evolution over decadal-to-millennial timescales (Istanbulluoglu and Bras, 2005). It seems more likely that short-term fluctuations in biological and physical processes are more important than any average activity (e.g. flood variability) (Reinhardt et al., 2010). Moreover, recent theoretical results from physical and numerical experiments also point to the important role played by vegetation cover on landscape evolution over millennial timescales (Istanbulluoglu and Bras, 2005; Tal et al., 2007).

Erosion is one of the most critical terrestrial environmental problems and the primary reason for soil degradation (Pimentel & Kounang 1998). It has been hypothesized that a vegetation cover with a high number of plant species of different growth forms, functional groups, and root characteristics will be more effective at governing interrill erosion processes than a vegetation cover with few species and a lower number of different growth forms (Körner & Spehn 2002). The assumed positive effect of plant diversity on landscape stability is still poorly supported by data, especially with regard to alpine ecosystems (Pohl et al.
Recent contributions emphasize the need to better quantify biotic processes and their contribution to landscape modification (Dietrich & Perron 2006; Amundson et al. 2007, Corenblit and Steiger, 2009; Richter & Mobley, 2009; Reinhardt et al., 2010). Current landscape evolution has the tendency to focus on the effect of a single species on geomorphic processes and one of the dominant views in the field of ecology is that every species has a unique ‘niche’ and imparts a unique signature on the physical environment (Dietrich & Perron, 2006). However, it is argued here that it is essential to consider the variance/diversity of life as a key landscape shaping process. The current evidence for the effect of plant functional types/diversity on key ecosystem functions derived from soil biological properties, such as litter decomposition, nutrient supply rates, erosion resistance, etc., is either contradictory or limited (Gastine et al., 2003; Zak et al., 2003; Wardle et al., 2004; Wardle, 2006). The effect of vegetation cover on the particle size of eroded sediment has mainly been described in relation to agricultural soils (Meyer et al., 1980; Deizman et al., 1987) and mining activities (Tordoff et al., 2000). The ability of wetlands to retain, and/or remove sediment and nutrients has long been recognized, particularly in the field of effluent disposal (e.g. Greenway and Woolley, 1999; Toet et al., 2005). It has been shown that vegetation cover can be effective in providing the necessary surface stability to prevent wind-blow of contaminated particulates from mining activities, and in reducing water pollution by interception of a substantial proportion of incident precipitation (Tordoff et al., 2000). However, little attention has been paid to the characteristics of the sediment eroded from the surface of undisturbed ground surface under natural vegetation (Martinez-Mena et al. 2000).

Some plants or plant communities are characteristic for stable conditions in a landscape, while others act as pioneers in recent deposits. A vegetation cover is very effective in reducing surface erosion because the roots bind the substrate. The development of the plant cover (either single-specimen or the community as a whole) can also be indicative of geomorphic process type and intensity (Goudie et al. 1990). Evidence supports that it is the degree of functional differences between species that is a driver of ecosystem processes, rather than the number of plant species (Bardgett, 2005). A study by Martin et al., (2010), presents the first evidence for a direct relationship between the diversity of plant functional groups and sediment yield in an alpine environment. No single species was correlated with sediment; only the combination of different species and growth forms explained a positive diversity effect. According to Liu et al., (2008) and Zak et al., (2003) the positive correlation between plant diversity and biomass production leads to an increase in the soil microbial biomass as enhanced biological activity can have positive effects on the formation of stable soil aggregates (Imeson & Vis 1982; Imeson & Verstraten 1989). Gitay & Nobel (1997) question whether consistent ecologically interpretable groups actually exist, and doubt that a universal functional classification can be defined because of the lack of agreement on traits in space and/or time. Nonetheless, the concept of plant functional types has emerged as a useful tool to provide balance between abstraction and detail to address large-scale ecosystem processes (Ustin & Gamon, 2010) as plant functional types (PFTs) at regional and global scales provide a coarse classification to predict vegetation dynamics (Prentice et al., 1992; Box, 1996; Leemans, 1997).

1.1.2 A multi-proxy analysis of palaeoenvironmental records

Dearing (2008) states that palaeoenvironmental records of ecological states and processes derived from geomorphic and limnic sediment sequences, together with information about climatic change, provide a wide range of information about the functioning of system behaviour (e.g. Dearing & Zolitschka, 1999). The study
of lake sediments can be directed towards reconstructions of the aquatic environment and/or of the terrestrial catchment of a lake, even including the regional landscape beyond the lake catchment (Birks & Birks, 2006). However, palaeoecological reconstructions based on a single proxy are limited (Ficken et al., 2002). Currently in palaeolimnological studies, a multi-proxy approach is the norm, and the aims of investigating ecosystem dynamics have turned more towards the reconstruction of past environments and climate changes (Lotter 2003). It is important to study as many proxies as possible in order to gain a wider overview of the complex network of interactions in an ecosystem than could be acquired from a single proxy (Smol, 2002; NRC, 2005). Every proxy has its own strengths and weaknesses, depending on such characteristics as its occurrence and abundance, temporal and spatial resolution, reaction and recovery time, and the climatic variables it approximates (e.g. Guiot, 1991). Extensive and detailed reviews of multi-proxy studies in palaeolimnology and palaeoecology include Wright (1966), Birks and Birks (1980), Delcourt and Delcourt (1991), Smol (2002), Cohen (2003), Lotter (2003), Pienitz et al. (2004) and NRC (2005). By combining proxies in any type of reconstruction study, strengths can be exploited and weaknesses can be identified (Mann, 2002).

A lake bed is the depositional site of both organic and minerogenic matter that is transported to the lake from the surrounding drainage basin, along with the organic matter that is produced/formed within the lake itself. The study of the amount of organic matter and grain-size measurements within a sedimentary profile is a widely used technique in palaeoenvironmental studies (Halfman and Thomas, 1984; Yu et al., 1990; Meyers & Vergès, 1999; Nesje & Dahl, 2001; Noren et al., 2002; Frangipane et al., 2009). The identification of the sources of organic and inorganic matter in a sedimentary record can provide a wealth of important palaeolimnological information on a lake and alluvial input changes through time, while the grain size distribution of sedimentary deposits is widely used as an indicator of sediment availability, flow competency and transport mechanism (Beierle et al., 2002). The choice of proxies for a palaeoenvironmental reconstruction is paramount. The proxies should be sensitive climatic indicators (i.e. their variations in the fossil record should be attributable to climate change) and complementary in their ability to reconstruct climate. Fast-reacting proxies such as e.g. stable isotopes, insects, or aquatic organisms may help to extract the climatic signal directly, whereas long-lived proxies such as upland vegetation and especially trees may show delayed reactions (Ammann, 1989a). The value of any multi-proxy study clearly rests on the reliability of the proxies used to reconstruct the past environmental conditions; different proxies reflect different environmental factors at a range of spatial scales and consequently show different strengths and weaknesses (Birks and Birks, 2006). The combination of such proxies may help to register leads and lags in the reaction of different biotic and abiotic systems to climatic change and thus enhance our ‘understanding of community ecology under a changing climate, including equilibrium or disequilibrium between communities and climate’ (Ammann, 1989a).

A functional plant cover provides soil stability and is especially important in alpine ecosystems (Pohl et al., 2009). However, there is currently a gap in our understanding of the erosion responses of mountain ecosystems. Few authors have taken an interest in the relationships between plant morphology and the effects they can have on catchment erosion. Therefore one of the primary aims of this study was to adopt a multi-proxy approach of reconstructing the palaeoenvironmental conditions at a carefully selected catchment employing the use of a novel method of using $^{10}$Be measurements to reconstruct past erosion rates in a high-relief setting. The Earth’s surface is constantly being exposed to a stream of particles arising from
nuclear processes that take place within our galaxy. Cosmogenic nuclides (i.e. $^{10}$Be) enable direct measurement of dates and rates of Earth’s surface processes as they build up predictably with time in minerals exposed to cosmic rays, therefore measuring their concentrations allows determination of how long rocks or sediment have been exposed at or near the surface of the Earth (Lal 1991; Gosse & Phillips 2001). Terrestrial cosmogenic nuclides (TCNs) are produced in-situ in bedrock and regolith by the interaction of cosmic rays, primarily nucleons and muons with target elements such as oxygen (Lal, 1988). They provide the first widely-applicable method of measuring surface processes over hundreds to millions of years. The development of techniques employing TCNs in both in-situ bedrock and sediment samples has revolutionised landscape denudation studies. Although the term ‘erosion’ is often used to describe the removal of mass from the Earth’s surface, the cosmogenic nuclide concentrations also reflect the loss of mass owing to chemical weathering (Strobl et al., 2012). In other words, concentrations of cosmogenic nuclides record the sum of physical erosion and chemical weathering, which is collectively referred to as denudation (Riebe et al., 2003; von Blanckenburg, 2005).

The importance of cosmogenic isotope analysis as a technique for this study arises from the timescales that it can address. Depending on the local rate of land surface stripping, cosmogenic isotope analysis can provide information on ages of geomorphic events, denudation rates and the operation of specific geomorphic processes over timescales ranging from hundreds to thousands of years; thus it can form a crucial bridge between investigations over the short term based on modern process rate measurements and historical data (Cockburn and Summerfield, 2004), and long-term studies using $^{10}$Be analysis. Cosmogenic $^{10}$Be measurements in river-borne quartz sand record a time-integrated erosion rate representative of an entire drainage basin and when sequestered in a deposit of known age, palaeo-erosion rates may be recovered from the nuclide content of the terrace material (Bierman & Steig, 1996; Granger et al., 1996; Schaller et al., 2002). Thus, this technique can be used to directly measure $^{10}$Be concentrations that are representative of mean catchment wide rates at the time of burial. The application of cosmogenic erosion rates in a depositional lacustrine setting is still relatively novel. It is used in this study to calculate palaeo-erosion rates in a high-relief catchment setting which is introduced in the following section.

1.2 Geographical context

1.2.1 Site Selection Sichuan

Basic research on ecological responses of high-elevation species to climatic variables is generally lacking in the Greater Himalayas (Figure 1.2). It is often too difficult to disentangle natural variability from potentially significant trends in data records which have been measured on a short timescales e.g. a few decades (Willis & Birks, 2006). Palaeo-ecology primarily considers the past, but it can provide a historical perspective to the present (National Research Council, 2005). One of the major research needs in conservation biology at the moment is an increase in palaeoecological studies of biodiversity hotspots- at present there are few studies from these critical areas (Willis & Birks, 2006). Further insight into the issues raised here can be achieved by analysing palaeo-records spanning the transition from glacial to interglacial conditions such as the Pleistocene- Holocene (P-H) boundary which enable inferences to be made about the effects of palaeo-monsoon behaviour on biodiversity and landscape dynamics (Hodell et al., 1999). One of the most important forces that has shaped the contemporary floristic richness of China is the climatic change that occurred
during the Late Quaternary period (López-Pujol et al., 2006). A review of available literature on Holocene climate change in Southwest China is discussed in section 1.2.4.

Figure 1.2 The Greater Himalayan region (red = alpine, all areas above 3000 m asl; yellow = montane, elevations falling between 1000 and 3000 m asl; green = lowland, elevation below 1000 m asl), taken from Xu et al. (2009).

The mountain regions of China (Figure 1.3) provide rich habitats for much of the country’s remaining biodiversity, owing to the heterogeneity of climates and soils, rapid elevational changes, varying aspects of slope direction, abundant microhabitats, and limited suitability for cultivation (Körner and Spehn, 2002). Global change is always reflected earlier in higher altitude regions before being transmitted to gradually to lower-altitude areas (Liu et al., 1997; Yao et al., 2000; He et al., 2004). The extremes of topography, vegetation and climate have served as a stimulus to evolution, and are responsible for the high species diversity and endemism of plants found in the Sichuan region, Southwest China (Sichuan Alpine Ecology Study Centre, 2009). These mountain regions still harbour large numbers of species, because there have been fewer human and natural disturbances and there are more diverse habitats (Tang et al., 2006). Thus, these mountains are likely to be especially important for preserving its remaining biodiversity (Chen 1998).
Montane vegetation has been shown to be particularly sensitive to climatic change (Birks & Ammann, 2000; Wick 2000) and small mountain lakes are regarded as reacting sensitively to environmental and climatic change (Mischke et al., 2005). Biota and soils along elevational gradients reflect the outcome of multiple interacting environmental factors over long periods of time (Becker et al., 2007). They are more likely to approximate steady-state responses to such conditions than experimentally manipulated systems. Hence, they have frequently been referred to as model test systems (Garten et al 1999) or experiments by nature (Körner 2003). Most of the services provided by mountain ecosystems, such as protection of slopes by vegetation cover and the productivity of forests and agro-ecosystems, are driven by the intimate linkages between plants, soils, and associated soil biota (Bardgett et al 2002). Despite the recent efforts to study mountain plant diversity in China, there are still significant gaps, especially in the remote western ranges (Li et al 2011). Until the 1980s, botanical exploration in the Sichuan region was scarce, and comprehensive inventories of the flora did not appear for nearly two decades (e.g. Kelley 2001). Mountain biodiversity is at present at high risk because of intensification and changes of land use (departing from traditional management), global climate change (shrinkage of glaciers), or the introduction of non-native species (Huber et al 2005; Spehn et al 2010).

China harbours probably the largest “evolutionary front” of the world’s North Temperate Zone, the Hengduan Mountains, which are still active because the Tibetan Plateau is uplifting (Zhang et al 2000) and which have experienced extensive explosive radiations in the last million years (Liu and Tian 2007). There is a distinct lack of palaeo-record coverage for the Hengduan Mountains and surrounding areas including the south eastern edge of the Tibetan Plateau (Zhang & Mischke, 2009). This region is currently one of the lesser-populated regions in China and is also one of the most inaccessible which limits the a priori potential impact of human activities (Lopez-Pujol 2011). Despite their inaccessibility, these mountains have been severely logged and overgrazed, and numerous dams have been planned or are already in place (CI 2007; Morell 2008), thus highlighting the importance of studying these ecosystems before human activity results in the significant modification of these natural environments.
1.2.2 Hengduan Mountain region

The Hengduan Mountain region in Sichuan Province, Southwestern China, has been designated one of the world’s biodiversity hotspots (Boufford and Dijk 1999, Myers et al. 2000, Boufford et al. 2004). The Hengduan Mountains include western and northwestern Yunnan, western Sichuan, eastern and southeastern Tibet (Xizang), southeastern Qinghai and southwestern Gansu, an area of 364 000 km² stretching from 24°40’ to 34°00’N and from 96°20’ to 104°30’E (Li, 1987) (Figure 1.4). In the north, the vegetation is less varied than in the south and is represented by the Qinghai- Xizang plateau alpine vegetation consisting mainly of coniferous trees (Liu et al. 1985, Yu et al. 1990). The mountains are at the margins of several plates, the Eurasian Plate to the north, the Indochina Block to the south, and the Indian Plate to the west (Chaplin, 2005). These plates are constrained by the Philippine-Pacific Plates to the east, and the Australasian Plate to the south (Hall 1997). This area which is situated along the eastern margin of the Tibetan Plateau lies on the Xianshuihe Fault zone, an elevated and sparsely populated region which ranges between 1500-4500m above sea-level. These differences in altitudinal gradient across the region have resulted in a marked climatic gradient with a range of geomorphological settings across the region, which contribute to the great species richness and diversity that are found there (Walker 1986; Li & Walker 1986). The Xianshuihe Fault (a strike-slip fault) extends for 350km in a northwest-southeast direction and is part of a 1400 km left-lateral fault system from southern Yunnan Province northwest through Sichuan Province into Qinghai Province (Xu & Kamp 2000). The varied geology, topography and climate have combined to create this epicentre of biodiversity that the Chinese Academy of Sciences considers the most critical region for biodiversity conservation in the country (IUCN 2003). Thus, it is a critical location for answering fundamental questions about the behaviour of the lithosphere and the processes that at upon its surface.

The Hengduan Mountains are undoubtedly the richest area for plant diversity in China but also one of the most significant (if not the most significant) of the North Temperate Zone (López-Pujol et al., 2011). The extremes of topography, vegetation and climate have served as a stimulus to evolution, and are responsible for the high species diversity and endemism of plants found in the Hengduan mountain region (SEPA 1998), although little research has been carried out on the changes in vegetation and landscape in this region since the last deglaciation.
1.2.3 Mountain vegetation Sichuan

Sichuan is the second-ranking Chinese province (including the Chongqing Municipality) in terms of plant biodiversity after Yunnan, containing about 9,500 species of vascular plants from over 1,600 genera distributed in 230 families (Lopez-Pujol et al., 2006). Nationally, the southern and northern sub-regions of the Hengduan Mountains have been recognized as being among the 11 key regions of terrestrial biodiversity in China (SEPA 1998). Until now, however, these areas have been poorly described, and their delimitation has been insufficiently analyzed (Zhang et al., 2009). Although this area represents only 5% of China’s total land mass, it harbours nearly half the total number (12,000) of all Chinese flowering plants, of which 3,500 are endemic to this area (Kelley 2001). Forest cover in Sichuan Province, one of China’s main sources of timber, declined from 19% in the 1950s to 12.6% in 1988 when a commercial logging ban was introduced (Conservation International, 2009). Forests in Sichuan are still being logged at a significant rate for firewood collection and house construction, particularly in the high valleys of the region (Conservation International, 2009).

The mountain regions still harbour large numbers of species in this region, because there have been fewer human and natural disturbances and there are more diverse habitats (Tang et al., 2006). Thus, the Hengduan mountains are likely to be especially important for preserving China’s remaining biodiversity (Chen 1998). The strong altitudinal climatic gradients that characterise mountain areas result in a steep ecological gradient which so there can be several ecotones occurring in a small area (Ortu et al., 2006). The area’s complex uplift history has fostered greater genetic diversity in the region because of complex patterns of exchange, isolation, adaptation, extinctions, and speciation (Chaplin, 2005). This can result in an
amplification of global climate signal (Beniston et al., 1997), therefore, vegetation response to climatic change is more pronounced at higher altitudes than in lowlands (Birks & Ammann, 2000). From a short term perspective, alpine areas are very fragile ecosystems and are highly sensitive to changing environmental conditions (Caviezel et al., 2013). It is poorly understood how sensitive the vegetation in this area is to climatic change and how it has responded to major changes in the past as the Holocene climate and vegetation history of this area in central Sichuan is still poorly investigated and the number of high-resolution records is limited.

1.2.4 Sichuan climate- what is known

In the Early Tertiary, the northward drift of the India-Australian Plate and its collision with the Eurasia Plate caused the gradual uplift and formation of the Tibetan Plateau (Molnar, 1989). The formation and uplift of this plateau had a hugely significant effect on the climate of this region as it resulted in the creation of the monsoons in east and south east Asia (Flohn, 1968; Manabe & Terpstra, 1974; An et al., 2000, Liu & Yin, 2002). Mainly due to strong land–ocean thermal contrast and the dynamic and thermal effects of the Tibetan Plateau, East Asia has a well-developed monsoon climate system (Fu, 2003). The monsoon formation over Asia is mainly triggered by the thermal contrast between land and oceans that is enhanced by the high elevated Tibetan Plateau (TP) being directly exposed to insolation (Prell and Kutzbach, 1992; Ye and Wu, 1998). The Asian monsoon is comprised of two subsystems, the Indian (or South Asian) monsoon and the East Asian monsoon. These are roughly divided at ~ 105°East (Figure 1.5) but both are a dynamic component of the modern climate system in this region. The inherent seasonal changes of monsoon circulation (atmospheric circulation and precipitation) can affect the ocean system, leading to strong seasonality in current strength and direction, sea-surface temperatures (SST) and salinity patterns (Wang et al., 2005). For this reason, there have been a considerable number of studies dedicated to the history of China’s monsoon climates (e.g. Flohn, 1968; Manabe & Terpstra, 1974; An et al., 1999, Liu & Yin, 2002, Wang et al., 2005; He et al., 2004, Herzschuh, 2005).

![Figure 1.5 Circulation systems influencing Southwest China (Indian Monsoon, SE Asian Monsoon, Westerlies) and present-day limit of the summer monsoon (after Gao, 1962, cited in An, 2000).](image)

Few studies have been carried out in the Hengduan Mountains which are strongly exposed to the East Asian summer monsoon (EASM) (Fan et al., 2009). This region is strongly influenced by the EASM during June-
September, while during the winter season continental air masses (East Asian winter monsoon) of the central Asian high pressure cell dominate and lead to very cold and dry conditions with occasional surges of dry continental polar air from northwesterly directions (Böhner, 2006; Fan et al., 2009). 70 to 90% of the annual precipitation falls between May and September, when the temperature is above 0°C (He et al. 2004).

High resolution palaeo-climate records for the region point to distinct phases of marked climatic fluctuation in the intensity of the monsoon climate during the Late-glacial/Early Holocene (Jarvis and Clay-Poole, 1992), with the Holocene particularly characterised by several centennial scale arid phases, superimposed upon a longer-term decline in monsoonal strength (Herzschuh, 2006). The tendency of temperature and precipitation inferred from pollen records has revealed the regularity of evolution of the EASM regions (Dykoski et al., 2005; Wang et al., 2001, 2008). Oxygen isotope (δ¹⁸O) records of stalagmites from several caves in China (Figure 1.6, Table 1.1) have yielded well dated, high-resolution records of the timing and amplitude of changes in East Asian Summer Monsoon (EASM) precipitation during the last four glacial–interglacial cycles (e.g. Cheng et al., 2009; Dykoski et al., 2005; Wang et al., 2001, 2008; Yuan et al., 2004). Generally, these records revealed that the EASM follows Northern Hemisphere (NH) insolation patterns (Peterse et al., 2011). The history of the EASM is an alternation between dominance by the dry-cold winter and warm-humid summer monsoons (An, 2000). Both precipitation and effective moisture are considered important environmental parameters within the region of the EASM, and are closely related to changes in winter and summer monsoon intensity (An et al., 2000). Studies have shown that a strong intensification of the EASM occurred at the Pleistocene-Holocene transition at ~11.5ka BP (e.g. Zhou et al., 1996, Fleitmann et al., 2003, Yuan et al., 2004; Herzschuh, 2005). In many studies the early Holocene in Southwest China was identified as a relatively warm and moist period (Herzschuh, 2006) with a strong summer monsoon that was reconstructed (along with other indicators) by the extension of forests on the eastern Tibetan Plateau (Yan et al., 1999; Shen et al., 2005; Herzschuh et al., 2006a; Shen et al., 2006a). The Holocene (or climatic) optimum, is often regarded as the most significant event in the monsoon history of this region. It is considered as a period of maximum postglacial warmth (e.g. Winkler and Wang, 1993) and an important climatic event which saw milder and wetter conditions during the early- to mid- Holocene in Southwest China. The Holocene optimum is important because it produced a varied array of geologic records and it may serve as an important analogue for future climatic change in this region (An et al., 2000). Although integrations derived from different proxies yield contrasting estimates for the timing of the Holocene optimum, they generally overlap during the period of 9-6 ka BP, suggesting that the optimum occurred sometime during the early- to mid-Holocene in Southwest China (Zhang et al., 2011). The late Holocene is mostly considered as a dry and cold period (Morrill et al., 2003; Herzschuh, 2006). Since 3 ka BP particularly, most records from the areas dominated by the EASM show markedly lower effective moisture (Herzschuh, 2006). The precipitation histories indicated by δ¹⁸Ocar records from different sites within the EASM region show a consistent pattern of progressively decreasing precipitation since the early Holocene (Zhang et al., 2011).
Figure 1.6 $\delta^{18}O_{\text{carb}}$ records in the monsoon region of China taken from Zhang et al. (2011). Data sources are listed below Table 1.

### Table 1.1 Information on sites with $\delta^{18}O_{\text{carb}}$ records in the monsoon region of China taken from Zhang et al. (2011).

<table>
<thead>
<tr>
<th>No</th>
<th>Sites</th>
<th>Location</th>
<th>Altitude (m a.s.l.)</th>
<th>Area (km²)</th>
<th>MAP (mm)</th>
<th>MAT (°C)</th>
<th>Carbonate Phase</th>
<th>$\delta^{18}O$ Range (%)</th>
<th>$\delta^{18}O$ Amplitude (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Lake Xingyin</td>
<td>24°20’N, 102°47’E</td>
<td>1723</td>
<td>39</td>
<td>900</td>
<td>15.6</td>
<td>Bulk carbonate</td>
<td>-12.1 to -4.9</td>
<td>7.2</td>
</tr>
<tr>
<td>2</td>
<td>Lake Qilu</td>
<td>24°10’N, 102°45’E</td>
<td>1797</td>
<td>37</td>
<td>900</td>
<td>15.6</td>
<td>Bulk carbonate</td>
<td>-9.9 to -5.1</td>
<td>48</td>
</tr>
<tr>
<td>3</td>
<td>Ahung Co</td>
<td>31°37’N, 92°34’E</td>
<td>4600</td>
<td>3.6</td>
<td>450</td>
<td>0.0</td>
<td>Bulk carbonate</td>
<td>-13.0 to -9.0</td>
<td>4.0</td>
</tr>
<tr>
<td>4</td>
<td>Selin Co</td>
<td>31°46’N, 88°50’E</td>
<td>4520</td>
<td>1640</td>
<td>200-300</td>
<td>&lt;0.0</td>
<td>Bulk carbonate</td>
<td>-5.9 to -1.1</td>
<td>4.8</td>
</tr>
<tr>
<td>5</td>
<td>Lake Koucha</td>
<td>34°00’N, 97°14’E</td>
<td>4530</td>
<td>18</td>
<td>510</td>
<td>-48</td>
<td>Ostracod shell</td>
<td>-5.4 to -0.2</td>
<td>5.2</td>
</tr>
<tr>
<td>6</td>
<td>Dongzi Co</td>
<td>35°20’N, 98°20’E</td>
<td>4080</td>
<td>229</td>
<td>311</td>
<td>-39</td>
<td>Ostracod shell</td>
<td>-4.1 to -2.3</td>
<td>6.4</td>
</tr>
<tr>
<td>7</td>
<td>Lake Qinghai</td>
<td>37°19’N, 99°00’E</td>
<td>3154</td>
<td>478</td>
<td>360</td>
<td>-5.7</td>
<td>Ostracod shell</td>
<td>-3.4 to -4.0</td>
<td>7.4</td>
</tr>
<tr>
<td>8</td>
<td>Hetongzhan Nur</td>
<td>39°17’N, 109°02’E</td>
<td>1350</td>
<td>dry</td>
<td>300</td>
<td>7-9</td>
<td>Claystone</td>
<td>-6.2 to -0.1</td>
<td>6.1</td>
</tr>
<tr>
<td>9</td>
<td>Lake Hamoertai</td>
<td>38°16’N, 108°20’E</td>
<td>1380</td>
<td>dry</td>
<td>300</td>
<td>7-9</td>
<td>Claystone</td>
<td>-6.1 to -3.8</td>
<td>9.9</td>
</tr>
<tr>
<td>10</td>
<td>Hetongzhan Nur</td>
<td>39°10’N, 109°16’E</td>
<td>1273</td>
<td>26</td>
<td>300</td>
<td>7-9</td>
<td>Bulk carbonate</td>
<td>-6.1 to -2.4</td>
<td>3.7</td>
</tr>
</tbody>
</table>

Data Sources: 1e2: Hodell et al., 1999; 3: Morrill, 2004; Morrill et al, 2006; 4: Gu et al., 1993; Morinaga et al., 1993; Kashiwaysa et al., 1995; 5: Mischke et al., 2008; 6: Mischke et al., 2010; 7: Lister et al., 1991; Liu et al., 2007; 8: Chen et al., 1995; 9: Wei et al., 1995; 10: Guo et al., 2007. MAP = Mean annual precipitation; MAT = Mean annual temperature.
1.3 Research hypotheses, aims and objectives

Palaeolimnology forms the basis of this research. A new palaeolimnological record from Lake Mugecuo, a high-altitude lake in the Hengduan Mountain region is explored here to reconstruct regional environmental dynamics since the Early Holocene. The conservation of biodiversity in a changing climate is considered to warrant a long-term temporal perspective (Willis & Birks, 2006). The PFT concept was utilised in this study to determine the changes in plant functional diversity at Lake Mugecuo through time in order assess whether plant functional diversity mediates the effects of climate on lake catchment dynamics (including long-term erosion rates). The method of classifying vegetation into specific plant functional type groupings was adopted here as a means of classifying the pollen data which are similar in a given set of traits and are similar in their association to certain variables (Pillar & Sosinski, 2003). A sensitive mountain catchment is used in this study to investigate the links between long-term changes in the climate and changes in plant community dynamics, and the potential impacts of these interactions on catchment palaeo-erosion rates. This study adopts a multi-proxy approach which takes advantage of the strengths and identifies the weaknesses of each proxy utilised.

The primary elements of this research are as follows:

- The palaeoecological analysis of fossil pollen grains
- The analysis of the organic content of the lake sediment archive using loss-on-ignition, %LOI, %TOC, %TN, C/N ratios and δ13C measurements
- The analysis of the physical properties of the lake sediments based on particle size analysis
- The direct measurement of modern and palaeo catchment-wide erosion rates in both in-situ, and lake sediment samples using cosmogenic isotope analysis
- AMS 14C radiocarbon dating of bulk sediment

This study aims to infer a causal link between climate, vegetation and landscape interactions. The focus is to determine the effects of changes in plant functional diversity on landscape change as well as determining whether plant functional types mediate the effects of climate change on landscape change (including erosion rates). A ‘multi-proxy’ approach is essential if meaningful inferences are to be made between these variables. This study aims to focus on a timescale of the last ~12,500 years (i.e. the Holocene period). A long-term approach is implicit to palaeoenvironmental studies in order to reveal the past response of a landscape to disturbance and to reveal patterns of resilience (Gil-Romera et al., 2010). This is an approach which has already proved useful in determining ecosystem dynamics under disturbance regimes, both human and naturally driven (e.g. Birks, 1996; Godwin, 1956; Carrión, 2002; Virah-Sawmy et al., 2009; Willis et al., 2005; Willis and Birks, 2006; amongst many others) but such studies are still severely lacking in biodiversity hot-spots worldwide. This study focuses on the Holocene period because of: a) the ease of obtaining lake cores from this period (Birks & Birks, 1980); b) the absence of significant human activity in the area (Kramer, 2010); and c) previous studies have shown that the climate record for this region is characterised by significant changes in climate over millennial – centennial timescale during the early-late Holocene (12 – 0 ka BP), superimposed upon a longer-term decline in monsoonal strength (Jarvis, 1992; Herzschuh, 2006). Such changes are likely to have influenced plant community dynamics throughout the
Holocene resulting in changes in the composition of the plant communities (fossil pollen data are used to provide evidence of this) and catchment dynamics but these are yet to be explored. Glaciation was not significant in this area post-LGM (Strasky et al., 2009) so a diverse ecosystem was likely present by the beginning of the early Holocene. The availability of a speleothem record which shows no sign of kinetic or evaporative effects and can be reliably interpreted as a record of local rainfall composition and temperature (Wang et al., 2005; Dykoski et al., 2005) is used a reference record of changes in effective moisture patterns in the Holocene for Lake Mugecuo. The research aims were to:

- Identify and examine the key environmental, vegetational and climatic shifts which occurred in Southwest China during the Holocene using the Lake Mugecuo sediment archive
- To examine the influence of these climatic shifts on plant functional diversity in the Mugecuo catchment by utilising fossil pollen data as a representative of changes in plant community dynamics through time in an alpine environment
- Compare the findings of this research with other regional records in order to provide a deeper insight into climate dynamics in Southwestern China.
- To determine if there is there any evidence of a relationship between changes in the vegetation record (i.e. plant functional diversity) and any changes in denudation rates/ lake catchment dynamics in the lake over time
- To determine if there is a direct link between climate change and the lake deposition record.
- To determine whether factoring in vegetation improves our understanding of the link between climate change and the lake deposition record.

The research objectives were to:

1. Produce a literature review which underpins this research (Chapter 1)
2. Collect and examine suitable lake sediment cores in from an appropriate site in Sichuan Province, China (Chapter 2)
3. Use a multi-proxy approach to the analysis of the lake core sediments (Chapter 2)
4. Produce a robust chronology for the lake cores (Chapters 2 & 3)
5. To use an existing regional climate record dataset to reconstruct Holocene climate at the study site (Chapter 3)
6. Compare this reconstruction with other regional proxy records in order to assess any notable trends in the dataset and identify possible drivers of these trends (Chapter 4)
7. Use the findings of this study to reconstruct the Holocene environmental history of the Lake Mugecuo catchment (Chapter 5)
8. Determine whether plant functional diversity mediates the effects of climate on landscape dynamics in the Mugecuo catchment (Chapter 5)
9. Infer a causal link between changes in plant functional diversity and lake catchment dynamics over time (Chapter 5)

10. To discuss the results of each of the proxies used in this study and present a critical analysis of the applicability and validity of this approach and its findings to the wider Sichuan region and similar catchment systems (Chapter 6)

1.4 Thesis outline/structure

This chapter has outlined the rationale for this research and provided a context for the research questions that are addressed. Chapter 2 provides a rationale behind the study site selection and detailed methodologies of the field, laboratory and statistical methods employed in this study. Chapter 3 presents all palaeoecological, palaeolimnological, palaeoenvironmental and palaeoclimatic results developed during this study. This chapter also presents chronological results and final age-depth models. Chapter 4 presents a detailed discussion on the palaeoecological and inferred palaeoclimatic reconstruction with a particular focus on the methodological evaluation of the plant functional diversity classification methodology adopted for this study. A synthesis of relevant published research in the context of palaeoclimatic reconstructions in this region is extensively discussed. Chapter 5 presents a detailed discussion of the palaeolimnological and palaeoenvironmental reconstructions of Lake Mugecuo with a particular focus on the determination of whether vegetation mediates the effects of climatic change on lake catchment dynamics based on statistical analysis of the data. Finally, Chapter 6 presents an integrated summary of all of the findings of the research and summarises the conclusions that may be drawn from this study, together with suggestions for the future direction of this research.

1.5 Project Partners:

This PhD research was part of a partnership that focuses on Late Quaternary climate and environment in Southwest China. The University of Exeter is a joint partner in several projects involving researchers based at the Nanjing Institute of Geography and Limnology (NIGL) and Professor Tong at the Chinese Academy of Geological Science (CAS) in China. The research aims of this partnership are to explore the relationships between climate and environmental change in Sichuan Province, China, based upon evidence derived from lake sediments. The project partners at NIGL have a Kullenberg Uwitech Coring Platform System which allows for easier sediment sampling in deep and shallow water. They have used this platform in many similar high-altitude lakes in the region so they have extensive field experience in this field area. As this was a collaborative effort, some of the data utilised by the author in this study was secondary data. However, the author was responsible for the sampling design and collection of the data.

The original aim of this research was to quantify the impact of climate change on the physical landscape as mediated by interactions between vegetation cover and catchment erosion rates. Unfortunately, due to unforeseen restrictions of the time we could spend working in the Lake Mugecuo National Park, there was insufficient time to collect sufficient material needed to fulfill the original research aim in terms of time we could spend coring the lake to obtain a long-term record of palaeo-erosion rates and surveying the modern catchment.
Chapter 2: Materials & Methods

Several lake sediment cores extracted from Lake Mugecuo which span the Holocene period, are used to assess the long term changes in environmental conditions, palaeo-vegetation and lake catchment dynamics in this region. Sichuan Province, Southwest China, presents an ideal location for this research due to the wealth of high altitude lake-catchment sites and existing palaeoclimatic information for the Late-Glacial/Holocene period (Jarvis, 1993; Wang et al., 2001, 2008; Yuan et al., 2004; Dykoski et al., 2005; Herzschuh, 2006). Due to the remoteness of the study site, human activity since the early Holocene has likely been minimal and thus it was assumed that anthropogenic factors were not a significant influencing factor upon the Lake Mugecuo record during the Holocene.

2.1 Site Description

Lake Mugecuo (30°9.200’N 101°50.921’E) is a large fresh-water body situated approximately 30km from the city of Kangding, the capital of the Tibetan Prefecture in western Sichuan Province. The lake (Figure 2.1) lies at an altitude of c.3800m a.s.l. and covers an area of c. 1.8km$^2$, with a maximum water depth of c.30m (Figure 2.1). The catchment covers an area of approximately c. 45km$^2$.

Figure 2.1 Map of the Lake Mugecuo catchment (30°9.200’N 101°50.921’E) in the Daxue Mountains in western Sichuan Province, China. Inset is a map of the lake bathymetry for reference.
Lake Mugecuo lies in the Daxue Mountains which are part of the Hengduan Mountains (Figure 2.3), making up the eastern margin of the Qinghai-Tibetan Plateau. It marks the transition zone between the low South Chinese plains and the Tibetan highlands, a rapid transition from a subtropical lowland climate to a high-relief mountain climate (Thomas, 1999). Thus, this region contains both subtropical and temperate vegetation elements, with several hundred endemic species. The Hengduan Mountain region, being at the plate margins, is an active earthquake zone (Chaplin, 2005). The lithology of the area is primarily made up of monzonitic granite and plagiogranite (Xu & Kamp 2000). Thermochronological studies suggest that the mountainous region in which the lake is situated (Daxue Mountains), must have been elevated since at least the Miocene (Kirby et al., 2002) and since this time, rivers have incised into the plateau and removed large volumes of sedimentary material (Kirby et al 2003). The Selaha-Kangding fault (60km long) is the main part of the Xianshuihe fault zone and extends from Jinglongsi, Selaha, Mugecuo to Kangding. Along the fault, there is evidence of recent movement, such as fault scarps and accompanying sag ponds with apparent horizontal offsets (Zhou et al., 2001). The Xianshuihe fault releases energy frequently through strong earthquakes due to high left-lateral slip rates and rapid energy accumulation (Cheng et al., 2011). Studies on the northern to central parts of the Xianshuihe fault (Wen et al., 1988; Zhao et al., 1992; Xu et al., 2003; Papadimitriou et al., 2004) have provided an estimated slip rate of 12 to 16 mm yr\(^{-1}\) (He & Oguchi, 2008). According to historical records, the Selaha-Kangding fault along which Lake Mugecuo lies produced a magnitude 7 earthquake in AD 1725, and it has been inferred that this fault has an interval of strong earthquake recurrence of ~300-400 yrs (Zhou et al., 2001).

Figure 2.2 Overview map of the eastern Hengduan Mountains showing major rivers and cities. The geographic setting is shown inset at the lower right. In the lower left is a DEM of the Kangding area, showing Zheduo Valley. The light dotted line marks the Sichuan-Tibet highway. The dark broken lines represent the five segments of the Xianshuihe fault zone (after Allen et al., 1991): X = restricted Xianshuihe fault, Y = Yalahe fault, S = Selaha (Kangding) fault, Z = Zheduotang fault, M = Moxi fault (taken from Strasky et al., 2009). The position of Lake Mugecuo is marked by a star symbol.
Lake Mugecuo is surrounded by slopes of varying steepness (Figure 2.3). The northern and eastern slopes are much steeper than the western and southern slopes. The relief in height from the observed water line to the highest ground level was estimated to be approximately 1,400m, 1,200m, 479m, and 1m for the northern, eastern, southern and western slopes respectively. The northern banks of the lake are characterised by scree slopes, high slope angles, sparse vegetation and frequent shallow landsliding (Figures 2.3-2.5). There is evidence of at least two substantial mass-movement events on the south western banks of the lake (Figure 2.5). These features consist of large boulders which may have been the result of peri-glacial action which transported and deposited debris from the valley during the past.

Figure 2.3 A 3D Google earth image of the Lake Mugecuo catchment. The white lines (numbered 1-7) provide examples of the varying elevation profiles of the catchment as indicated by the corresponding panels on the right. The blue arrows represent the likely direction of movement of sediment through the catchment.
Climatic conditions in the region are strongly related to elevation yielding steep environmental gradients (Kramer, 2010). The climate of the region is relatively warm and humid (mean annual temperature is 7.8°C) and the area is strongly influenced by the southwest EAM which provides most of the annual precipitation of typically more than 1000mm (Shi, 2002). The temperature ranges from 5 to 21 °C for the warmest month and precipitation rates vary from 500 mm to ~1200 mm in the region (Yu et al., 2001). The nearest climate station is Garzê (31.62° N, 100.00° E, 3522m a.s.l.) where 90% of the annual precipitation falls between May to October and the mean annual temperature (MAT) is 6.6 °C and mean July temperature (MJT) is 14.3 °C. Based on the observed temperature gradient of 0.5 °C decrease per 100 m increase in altitude (Böhner, 1994), the estimated MJT is 15.8 °C and estimated MAT is 8.1 °C at Lake Mugecuo. A review of palaeoecological studies for Southwestern China points to distinct phases of marked climatic fluctuation in the intensity of the monsoon climate during the Late-glacial/Early Holocene (Jarvis and Clay-Poole, 1992).
with the Holocene particularly characterised by several centennial scale arid phases. These phases were superimposed upon a longer-term decline in monsoonal strength (Herzschuh, 2006).

Modern catchment vegetation primarily consists of montane (Figure 2.6) forest up to around 4,000m composed of conifers such as *Pinus* and broadleaves such as *Betula*, along with many drought-tolerant species such as *Artemisia* which tolerate the monsoonal climate which is characterised by cold, dry winters and wetter, warmer summer temperatures (~7.8°C). The western banks of Lake Mugecuo are currently dominated by alpine meadow/scrub vegetation along with a prominent floodplain (Figure 2.7), while the southern and eastern banks of the lake are dominated by closed forest (Figure 2.6).

![Figure 2.6 Montane forest cover in the Lake Mugecuo catchment.](image)

![Figure 2.7 Modern alpine meadow/grassland on the valley floor/floodplain on the western banks of the lake.](image)

### 2.2 Reasons for site selection

This site was selected because the granite lithology of the catchment is amenable to $^{10}$Be erosion rate measurement and the need for high (temporal) resolution palaeo-erosion data. The elevation range between the summit and the valley floor in the Lake Mugecuo catchment is >1,000m (Figure 2.8). The high relief and steep slopes which are the distinguishing characteristics of mountain terrain provide much potential energy for erosion and sediment transport (Figure 2.9). This potential is translated in turn into the rapid erosion rates which have long been recognized as typical of mountain regions i.e. erosion rates increase with increasing
slope gradients (Fournier, 1960; Schumm, 1963; Young, 1969; Bierman & Caffee, 2001; Codilean, 2007). Therefore it is likely that evidence of past denudation change would be evident in the Mugecuo sediment archive due to the high relief and topography of the catchment.

Figure 2.8 3D image of Mugecuo catchment area (90m raster DEM). The elevation range between the summit and the valley floor in the Lake Mugecuo catchment is >1,000m.

Figure 2.9 Plot of $^{10}$Be denudation rates versus mean catchment slope from cosmogenic nuclide data of Bierman & Caffee (2001) and Codilean et al., (2007).

In-situ bedrock samples record bedrock exposure age or erosion rate (Lal, 1991), and the measurement of $^{10}$Be concentrations in river sediments enables calculation of catchment-wide (detrital) erosion rates (Bierman and Steig, 1996). In-situ erosion rates can be calculated from $^{10}$Be concentrations in bedrock, and mean catchment erosion rates calculated from the $^{10}$Be content of detrital sediment provided that erosion is steady, and that catchment-wide $^{10}$Be production is balanced by $^{10}$Be leaving the catchment (i.e. there is no significant sediment storage) and that the quartz is uniformly distributed within the source area e.g. in a
granite-dominated catchment (Lal 1991; Brown et al. 1995; Bierman and Steig 1996; Granger, Kirchner et al. 1996; Stone 2000). Long-term erosion rates inferred from cosmogenic nuclides are insensitive to changes in short-term (<100 yr) fluvial sediment storage (Granger et al. 1996), particularly in small, steep catchments such as the Lake Mugecuo catchment (average hillslope gradients >0.10), where sediment storage area is typically small relative to total catchment area. Cosmogenic nuclide erosion rate measurements are integrated over the time required to erode ~60cm of material: at an erosion rate of 1mm/yr the $^{10}$Be integration time is 600 years (Bierman & Nichols, 2004). Thus, high temporal resolution data can only be obtained in high-relief rapidly eroding terrain with relatively short integration times such as the Lake Mugecuo catchment. Significant periods of partial or complete shielding of the landscape from cosmic rays by glaciers can hamper the process of calculating catchment-wide erosion rates, as such calculations will underestimate the minimum total exposure history (Nishiizumi et al., 1989). There is no evidence of Late-Pleistocene glaciation in the Mugecuo catchment. No modern glacier exists in the catchment area as it lies below the current snowline whilst contemporary glaciers only exist around the highest summits, and permanent snow occurs above 4500m a.s.l. (Strasky et al., 2009). Glaciation was not significant in this area post-LGM (Strasky et al., 2009) so a diverse ecosystem was likely present by the beginning of the early Holocene.

The availability of a speleothem record which could be used a reference record of changes in effective moisture patterns in the Holocene for Southwest China (Wang et al., 2005; Dykoski et al., 2005) made this a suitable site for this research. It has been previously outlined that the climate record for this region is characterised by significant changes in climate over millennial – centennial timescale during the early-late Holocene (Jarvis, 1993, Herzschuh, 2006). It was inferred that these major changes would likely be captured in the natural sediment archive at Lake Mugecuo. The absence of significant human activity in the area (Kramer, 2010) which limits the potential impact of human activities on the natural sediment archive also made this site an appropriate study site choice. The Mugecuo national park is now designated as a scenic area for the benefit of tourism development in this area (Schedler et al., 2010). It is one of the key regions of terrestrial biodiversity in China (SEPA 1998) and is thus an important biodiversity hotspot which preserves China’s remaining biodiversity. Lake Mugecuo is approximately 6.2 km in diameter and therefore the pollen deposited in the lake likely reflects regional and local vegetation and inferred climatic conditions (c.f. Jacobsen et al., 1981). As mentioned previously in Chapter 1, the project partners for this research have much experience in lake coring in this region making this an appropriate choice of study site for this collaborative research.

### 2.3 Lake Sediment Coring

#### 2.3.1 Core site selection

A deep-water core (MG1) forms the focus of the palaeoecological and the majority of the palaeolimnological work of this project with a combination of pollen analysis and lithological/geochemical techniques (loss-on-ignition, total organic carbon, total organic nitrogen, C/N, $^{13}$C, particle size) employed to reconstruct past catchment vegetation dynamics and past variation in sediment dynamics within the lake. A history of catchment erosion rates is reconstructed here using cosmogenic $^{10}$Be concentration measurements of river-borne quartz sand. Two shallow water lake cores (MG3 & MG4) provide an independent measure of
catchment-wide erosion rates due to the coarse sandy material required for this analysis (see Figure 2.10 for project schematic and section 2.6 for cosmogenic isotope analysis methodology).

The width of the floodplain at the northern end of the lake suggests that the channel has meandered across the valley over time. Therefore, the sandy (deltaic) cores collected from this site were perfect for the $^{10}$Be analysis proposed here as there is evidence in the lake core sediments of channel avulsions, in the form of several multi-centimetre sand layers interspaced with thicker layers of mud (see Figure 2.11). The purpose of utilising these channel avulsions for palaeo-erosion rate measurements is explained in section 2.6.2.

The deep water core is very fine-grained and organic-rich with plentiful pollen while the two shallow water cores were low in organics, rich in sand and contained negligible quantities of pollen. Thus it is not possible to directly compare $^{10}$Be erosion rates with pollen concentrations in a single core. To draw such a
comparison three cores were required for this study; a deep water core rich in pollen but no sand and two shallow water cores rich in sand but lacking pollen. AMS $^{14}$C dating provided the mechanism for correlating between these disparate cores. By determining a robust chronology for each of these cores, the results of the multi-proxy analysis carried out on the deep water core could be directly compared over time periods that were synchronous with the mean catchment wide erosion rates (measured from $^{10}$Be concentrations from the sand layers) in the shallow water cores. The use of this relatively novel technique expands on the multi-proxy approach of this study in reconstructing the palaeoenvironmental dynamics of a high alpine lake catchment in the Hengduan Mountains, Southwest China as there is a current a paucity of such data for this region.

2.3.2 Extraction of cores from Lake Mugecuo

A total of three lake sediment cores were extracted from Lake Mugecuo (Figure 2.12) in July 2010 using a Kullenberg Uwitech Coring Platform System by scientists from the University of Exeter and NIGL (Figure 2,13). Details of these cores are outlined in the following sections.

![Isolines of water depths of Mugecuo Lake](image)

**Key**

- Shallow Water Coring Site
- Deep Water Coring Site

Figure 2.12 Lake Mugecuo bathymetry map with coring sites highlighted.
2.3.3 Deep water core MG1

A 3.38m long core (MG1) was extracted from the deepest part of Lake Mugecuo (30°9.053'N 101°51.378'E) where turbulence and mixing of the lake sediments is minimal. The water depth at this point is approximately 30m (Fig 2.12). A total of 3.4m of sediment was recovered from the lake in three sections. It was hopeful at this point that the record would cover the whole Holocene period or even as far back as the LGM. Prior to analysis and shipment back to the UK, the MG1 core was sub-sampled at 1cm intervals and kept in cold storage. The samples were then freeze-dried and stored in cold storage in the cold room at the University of Exeter, Cornwall Campus. In some sections of the MG1 core there was substantial doming of the sedimentary bands creating curved bell-shaped strata (see Figure 2.14). These deformations were up to 5cm in some cases. The doming was most likely related to frictional forces on the tube wall as it delved deeper into the compacted sediment. To overcome the problem of these distortions, a hollow (ITRAX) channel was placed in the middle of the coring tubes which meant that the core could be sub-sampled from the sediment in the channel as opposed to the distorted sediment either side of this channel. Distortion of the sediment in the centre of the channel was negligible thus the material from the ITRAX rail was used for the multi-proxy analysis of the MG1 core, thus minimising the impact of this doming effect on the data (the edge material was discarded). The top few cms of sediment were not significantly disturbed by water or during the core extraction process.

The deep water core (MG1) was extracted in three sections. Due to the difficulties in extracting the deep water core MG1, it was not practical to obtain running overlaps for the three sequences extracted so the assumption was made that there was no overlapping between them i.e. it was not necessary to remove any overlapping samples from the composite core record. Great care was taken during the core extraction process to prevent any loss of sediment which could compromise the composite MG1 core record.
2.3.4 Shallow water cores MG3 and MG4

Two shallow water cores MG3 (30°9.504 'N 101°50.157'E, measuring 4.64m) and MG4 (30°9.053 'N 101°50.167'E, measuring 3.12m) were recovered from the main river delta entering the north-western end of the lake at a water depth of ~5m in July 2010 (Figure 2.12). In each case coring continued until the coring tube could go no deeper. The shallow water cores MG3 and MG4 were also extracted in three sections. Again the assumption was made that no overlapping occurred, enabling a composite core profile to be used from the sequences extracted for each of the shallow water cores.

Prior to analysis and shipment back to the UK, cores MG3 and MG4 were subsampled at 1cm intervals and kept in cold storage. The samples were then freeze-dried and stored in cold storage in the cold room at the University of Exeter, Cornwall Campus. These cores were taken with the hope of finding material dating back to the early Holocene but preliminary dating (see section 3.2) revealed that the deepest material was ~3,800 cal.yr BP.

2.4 Core chronologies

2.4.1 AMS $^{14}$C radiocarbon dating

An absolute chronology is the basis of all comparisons and correlations of Late Quaternary stratigraphic proxy records (Telford et al., 2004). A chronology for each of the cores collected at Mugecuo Lake was obtained using Accelerator Mass Spectrometry (AMS) $^{14}$C dating of bulk sediment samples, a well-established technique for many Holocene studies (Brown et al., 1989; Shuman et al., 2002; Williams et al., 2004). The fundamentals of radiocarbon dating are discussed in detail by Lowe and Walker (1997) and Bronk Ramsey (2008a).

The benefit of radiocarbon $^{14}$C dating is that it reduces the amount of material needed for dating which allows for additional analysis on the remaining sediment. Each of the core chronologies in this study are primarily based on AMS radiocarbon dates of bulk sediment by the NERC Radiocarbon Laboratory and the Oxford Radiocarbon Accelerator Unit. This was due to a distinct lack of terrestrial plant macrofossils in the lake core.
sediments. After several unsuccessful attempts at AMS dating of charcoal and pollen, the decision was made to proceed with bulk sediment dating to obtain a robust chronology for each of the cores.

The selection of samples for dating in the deep water core (MG1) was based upon the composite core stratigraphy and the pollen record in order to obtain a robust chronology for the entire sediment sequence. The selection of material for AMS $^{14}$C dating in the shallow water cores was based on the need to constrain the ages of the bands of sand sediment that were used for $^{10}$Be erosion-rate measurements in the cores (see section 2.6 for further explanation of the palaeo-erosion samples). Several attempts were made to use a pollen dating technique but the sediment adjacent to the units of sand in the shallow water cores was simply too gritty and organic-poor to archive the required quantity of pollen necessary for $^{14}$C analysis. The selection of samples for dating in each of the cores was based upon the composite core stratigraphy, particle size data, and the availability of dateable material which are further discussed in section 3.2.

### 2.4.2 Calibration of $^{14}$C dates and the construction of $^{14}$C age models

Owing to past changes in atmospheric $^{14}$C concentration, the relationship between $^{14}$C and calendar age is far from linear, thus requiring a calibration curve (Reimer et al. 2009). To calibrate a $^{14}$C date, the aim is to work out the probability of that date belonging to a given calendar year. The current most widely accepted calibration curve is IntCal09 (northern hemisphere atmospheric; Reimer et al. 2009). All $^{14}$C dates in this study were calibrated and tested for internal consistency using Oxcal v. 4.17 and Intcal09 (Bronk Ramsey, 1995; Bronk Ramsey, 2008; Reimer et al., 2009). The calibration of $^{14}$C dates using Intcal09 is based upon an approach in which independent increments from one calendar year to the next are drawn from a Gaussian distribution. Markov chain Monte Carlo (MCMC) analysis is used to generate posterior realizations of the complete calibration curve simultaneously (Reimer et al., 2009). The modelled dates were calibrated to AD 1950 calendar years before present (cal. yr BP), with a > 95% probability that the dates fell within expected parameters as defined by a normal distribution curve based on known ages derived from dendrochronological and marine sources (Reimer et al., 2004).

Age-depth models have been designed to estimate the ages of depths in sedimentary deposits, based on limited numbers of dated depths and on assumptions as to how the deposit has accumulated between those dated depths over time (Blaauw, 2010). The number of dates needed to construct an age-depth model for a sediment sequence depends on the precision required and the complexity of the sedimentation rate of the deposit. Bayesian statistical methods have been shown to produce reliable age-depth models (e.g. Blaauw and Christen, 2005; Bronk and Ramsey, 2008) as it makes use of sophisticated iterative sampling methods. The Oxcal v.4.17 software uses a Bayesian approach which essentially samples over all of the possible ages for each depth point to give prior and likelihood probabilities; the resulting distributions known as ‘posterior probability densities’ take account of both the deposition model and the actual age measurements made. In this study, the Bayesian P_sequence depositional model was used in Oxcal v.4.17 to construct an age-depth model for each of the MG cores (Appendix A.1.1). The OxCal P-sequence allows for fluctuations in the deposition rate, assuming it to be mediated by the Poisson process with a given step size, represented by the model parameter ‘k’ (Goslar et al., 2009). The ‘k’ factor must be assigned to the model to specify and estimation for the number of events per given depth and the overall stratigraphical process (Blockley et al., 2007). The model assumes that a) as age increase with depth, radiocarbon dates can be modelled in a specific order and b) that fluctuations in the rate of deposition can occur (Bronk-Ramsey, 2008b).
2.4.3 Unified core chronologies

As outlined previously it was not possible to simply directly compare $^{10}$Be erosion rates calculated from two shallow water cores with pollen concentrations in the deep water core; to draw this comparison it was necessary to correlate the shallow water core chronologies with the deep water core chronology. $^{14}$C dating provided the mechanism for correlating between these disparate cores through the construction of a unified core chronology. It is important to stress that $^{10}$Be erosion rates give catchment wide erosion rates for specific periods of time that are unrelated to depositional environment. Thus once chronologies were established, erosion-rates measured in the shallow cores could be related equally to rates of sedimentation in both the shallow cores AND in the deep water core as each of the cores spanned concurrent periods.

2.4.4 Sedimentation rates

Mean (smoothed) sedimentation rates between dated points were calculated for each of the cores to a) gain an insight into the shifts in rates over time and b) to estimate the chronological resolution of the cores i.e. the number of years represented by each cm of sediment. The calculation of sedimentation rates is indicative only- rates may have changed between dated samples so smoothed calculated rates between them may have been more or less variable over shorter timescales.

As well as modelling the AMS $^{14}$C dates in the MG1 core, the OxCal P-sequence model was used to model sediment accumulation rates. The model assesses the degree to which sedimentation rates are uniform or more variable down the length of the core sequence. The best agreement was achieved by setting the k value to one (implying a variable sedimentation rate down the core). Sedimentation rates were estimated by dividing the number of years in the interval between each pair of dates by the thickness of the intervening sediment.

2.5 Organic and lithological analyses

2.5.1 Physical properties and organic analysis

The geochemical composition of lacustrine organic matter can provide information about the source and amount of organic material entering a lake system and thus provide information on vegetation changes through time (Meyers & Teranes, 2001). Techniques such as loss-on-ignition, total organic carbon, total organic nitrogen, C/N, $\delta^{13}$C and particle size analysis are utilised to provide such information in this study. Loss-on-ignition provides an insight into sediment composition and is a widely used method for estimating the organic and carbonate content of sediments (Dean, 1974; Bengtsson & Enell, 1986), while TN% and TOC% can be used as proxies for palaeo-productivity rates and the sources of organic material in the sediment archive (Shen Ji et al., 1997; Zhu et al., 2002; Cook et al., 2011). C/N ratios and $\delta^{13}$C used together provide a record of changes in organic matter composition through time (Lamb et al., 2004). Particle size analysis can provide an insight into catchment stability and erosion rates (Loveland & Whalley, 2000; Sperazza et al., 2004). Collectively, these variables provide a detailed picture of past environmental changes at Lake Mugecuo. All analyses were carried out on the same core to minimize errors of stratigraphic correlation as per Lotter (2003).
2.5.2 Loss-on-ignition

The primary source of organic matter to lake sediments is from the detritus of plants which have lived in the lake or its watershed. The loss-on-ignition (LOI) technique can be used to determine variations in sediment input to the basin and its influence on the proxy record. It is generally accepted that organic materialcombusts at approximately 550 °C whilst carbonate minerals tend to combust at ~925-950 °C (Santisteban et al., 2004). A standard LOI technique was used as per Heiri et al. (2001) which required the combustion of the sediments at a high temperature (550°C for 4 hours) to quantify the proportion of organic material within the lake sedimentary profile. The weight loss during the reactions is measured by weighing the samples before and after heating and is closely correlated to the organic matter and carbonate content of the sediment (Dean, 1974; Bengtsson & Enell, 1986).

Loss-on-ignition analysis was carried out on MG1 at a resolution of 5cm. This technique was used to estimate the percentage of organic and inorganic present in the sediment core based upon the relationship between combustion temperature and sediment type. It is a percentage, and thus an increase can reflect an absolute increase in organic matter or an absolute decrease in mineral matter, or some combination of both (Birks & Birks, 2006). Therefore, it is an indicative method only.

Following the techniques as per Heiri et al. (2001), the crucibles were dried at 105°C for 1 hour. They were then allowed to cool and were subsequently weighed (Wc). The crucibles were then half filled with the samples which were oven dried at approximately 105°C. The crucibles plus the samples were allowed to cool and then reweighed (Ws). The samples were then placed into a furnace for 4 hours at approximately 500°C. The crucibles plus the ash were then allowed to cool fully and were then weighed once more (Wa). The loss-on-ignition was then obtained using the following calculation:

\[
\frac{Ws - Wa}{Ws - Wc} \times 100\%
\]

2.5.3 Total organic carbon, total organic nitrogen and C/N

In aquatic systems, primary organic material is produced by photosynthetic production by phytoplankton. Most organic matter contains approximately 40% organic carbon (Leng, 2009). Total Organic Carbon (TOC%) specifically measures the organic carbon content of sediments and is a measure of lake productivity (Zhou et al., 2007). TOC% is a fundamental parameter for describing the abundance of organic matter in sediments. TOC concentrations are influenced by both initial biomass production and subsequent degree of degradation, so they integrate the different origins of organic matter, delivery routes, depositional processes, and consequent degrees of preservation (Meyers & Lallier-Vergès, 1999). Lake size may also influence nutrient availability. Larger lakes tend to be less productive than smaller lakes, as they receive less concentrated supplies of nutrients through overland flow and runoff from terrestrial surfaces (Meyers, 1997).

Due to being similar measures of lake productivity, the results of LOI and TOC% analysis should be very similar in terms of the proxy curves produced, but dissimilar in terms of the relative percentages recorded. LOI analysis was used to infer broad organic trends in the MG1 core. %TOC was then used to constrain the interpretation of organic carbon content in the core.
%TOC analysis was carried out at 2cm intervals on MG1 (the deep water core) in order to estimate the amount of organic carbon content of the sediments, expressed as a percentage. Percentage carbon was measured on sediments which were initially treated with 5% HCl (to remove carbonates) and were then measured using a Carlo Erba elemental analyser using the standard technique outlined in Leng et al., (2005) by staff at the Nanjing Institute of Geography and Limnology. The elemental analyser is based on a series of sequential steps. The samples were burnt and then the gases produced are flushed with pure helium through various stages to a detector. The output of these stages are then interfaced to a computer to give the final determinations.

Total Nitrogen (%TN) values reflect changes in aquatic nitrogen fixing and cycling, thus providing evidence about the source and palaeo-productivity of sediment organic matter. Nitrogen is derived from several sources including the atmosphere (as dissolved organic and inorganic nitrogen), and water systems such as streams and groundwater (as particulates and dissolved organic nitrogen originating from plants and soils and dissolved inorganic nitrogen as nitrate or ammonium) (Leng et al., 2006). TN analysis was carried out at 2cm intervals on MG1 (the deep water core). Percentage nitrogen was measured on sediments treated with 5% HCl (removes carbonates) using a Carlo Erba elemental analyser using the standard technique outlined in Leng et al., (2005) by staff at the Nanjing Institute of Geography and Limnology.

Lake water carbon and nitrogen balances are strongly influenced by changes in catchment hydrology, soil decomposition and vegetation (Wolfe et al., 1999). C/N ratios are a useful indication of the relative proportions of autochthonous (organic aquatic particles) and allochthonous (terrestrial organic particles) sources of lacustrine organic matter (Talbot & Laerdal, 2000). The C/N ratios of lake sediments may reflect the proportions of terrestrial versus algal carbon contributing to the accumulation of sediment (Kaushal & Binford 1999). As an accumulation of ‘geochemical fossils’, the organic matter content of lake sediments provides information that is important to interpretations of lacustrine palaeoenvironments, histories of climate change, and the effects of humans on local and regional ecosystems (Meyers & Lallier-Verges, 1999). C/N values were derived from the %TOC and %TN records (calculated by molecular weight). Aquatic vegetation predominantly has C/N ratios of 4-10, whereas terrestrial vascular plants generally have ratios >20 (Lamb et al., 2004). Ratios between 10-20 suggest a subequal mixture of algal and vascular plant contribution, which is expected for most lake systems (Meyers & Lallier-Vergès, 1999).

2.5.4 \( \delta^{13}C \)

\( \delta^{13}C \) analysis of bulk organic sedimentary matter is another tool used for palaeoenvironmental reconstruction. \( \delta^{13}C \) values from organic-rich lake sediments have been shown to reflect the \( \delta^{13}C \) composition of the plant material entering the lake (e.g. Olago et al., 2000). The \( \delta^{13}C \) value of organic matter in lacustrine sediments is generally influenced by the proportion of terrestrial, near shore aquatic plants, plankton and the chemical characteristics of the lake water (Shen Ji et al., 2005). \( \delta^{13}C \) analysis is based on the premise that plants fix carbon via one of two pathways (Meyers et al., 1999). C\(_3\) plants (those that survive solely on C\(_3\) fixation) typically fractionate CO\(_2\) values of approximately -27‰ and C\(_4\) plants (those that survive solely on C\(_4\) fixation) produce \( \delta^{13}C \) values of approximately -13‰, while CAM (Crassulacean Acid Metabolism) photosynthesizers produce a variety of fractionation effects which range from -4 to -20‰ (Lamb et al., 2004; O'Leary, 1988). Bulk organic \( \delta^{13}C \) analysis on its own is limited because it does not distinguish
between different sources of organic matter entering a lake (Lamb et al., 2004). Therefore it is often used in tandem with C/N ratios to identify the sources of material preserved in lake sediments. By plotting $\delta^{13}C$ values against C/N ratios (Figure 2.15), it is possible to detect changes in plant type abundances, lake productivity and carbon cycling (Leng, 2006). $\delta^{13}C$ analysis was carried out on the deep water core MG1 at a resolution of 2cm in order to make it comparable to the 2cm pollen record. $\delta^{13}C$ analysis was carried out using a Carlo Erba elemental analyser using the standard technique outlined in Leng et al., (2006) by staff at the Nanjing Institute of Geography and Limnology.

![Figure 2.15](image)

Figure 2.15 Representative elemental and carbon isotopic compositions of organic matter from lacustrine algae, C3 land plants, and C4 land plants that use CO$_2$ as their source of carbon during photosynthesis. Deviations from these generalized patterns occur and provide paleolimnologic information (Taken from Meyers & Lallier-Verges, 1999).

2.5.5 Particle size

The physical attributes of sediment grains such as particle size are often used in multi-proxy studies as they indirectly respond to changes in climate and/or limnology (Last et al., 2001). Changes in the size and distribution (sorting) of clastic lake deposits vary with catchment erosion, transport and deposition processes which are controlled by a variety of factors including precipitation rates and vegetation cover (Leeder et al., 1992). Particle size analysis can provide evidence of in-wash of sediment into a lake basin (Liu et al., 2008; An et al., 2011). It can also reveal changes in the proportions and types of sediment in a profile, caused by changes in environmental conditions or changes in the surrounding landscape. Grain-size distributions of naturally occurring sediment have long been used as an important source of information for the interpretation of sedimentation style (e.g., distinguishing underflow from suspension settle-out deposition) and environmental reconstruction (e.g., distinguishing littoral from offshore settings) (Sperazza et al, 2004). The physical energy of lake water is the most important factor controlling the particle size distribution of lake sediments (Chen & Wan, 1999). Fine-grained sediments and coarse-grained sediments respectively are said to represent high and low physical energy of lake water (Hakanson et al., 1983; Zhang Shi et al., 1997).

Particle size analysis measures the size (diameter) of clastic particles contained in a sediment sample. Particle grain size was measured at a resolution of 1cm on the deep water (MG1) core and at a 4cm
resolution on the shallow water cores (MG3 & MG4) using standard techniques outlined in Last et al. (2001). Bulk sub samples (~10g) were treated with 10% H$_2$O$_2$ and 0.4% hexametaphosphate solution in order to remove organic material. The remaining sediments were analysed using a laser particle-size analyser (Master Sizer 2000, Malvern Instruments Ltd) which measures particles ranging from 0.1 to 1000µm in size. Particle size (PS) measurements were returned in the form of five clastic particle sizes: PS < 4.00 µm, PS >4µm < 16µm, PS >16µm <32 µm, PS >32µm <64µm and PS >64 µm. The sediments were then categorised according to the Udden-Wentworth scale where sand is classified as >63 µm, silt 2-63 µm or clay <2 µm (Last et al., 2001; Leeder, 1992).

2.5.6  X-ray diffraction (XRD) mineralogical analysis

X-ray diffraction is a technique widely used for quantitative analysis of geological samples. It was used in this study to provide a detailed characterization of the clay bands recorded in the MG1 record (see Figure 3.1). X-ray diffraction (XRD) mineralogical analysis of the clay bands was carried out at the BGS laboratory at Keyworth, Nottingham. The <63 µm material was placed in a 250 ml measuring cylinder with 0.5 ml 0.1M sodium hexametaphosphate ('Calgon') solution to disperse the individual clay particles and prevent flocculation. After standing for a period determined using Stokes' Law, a nominal <2 µm ('clay') fraction was removed and dried at 40°C. Approximately 100 mg of the <2 µm material was re-suspended in a minimum of deionised water and pipetted onto a ceramic tile in a vacuum apparatus to produce an oriented mount. The mounts were Ca-saturated using 2 ml 0.1M CaCl$_2$.6H$_2$O solution and washed twice to remove excess reagent and allowed to dry at room temperature.

XRD analysis was carried out using a PANalytical X'Pert Pro series diffractometer equipped with a cobalt-target tube, X'Celerator detector and operated at 45kV and 40mA.

<2 µm oriented mounts were scanned from 2-40°2θ at 1.02°2θ/minute after air-drying, ethylene glycol solvation and heating at 550°C for 2 hours. Clay mineral species were then identified from their characteristic peak positions and their reaction to the diagnostic testing program.

In order to gain further information about the nature of the clay minerals present in the samples, modelling of the <2 µm glycol-solvated XRD profiles was carried out using Newmod-for-Window™ (Reynolds & Reynolds, 1996) software.

Modelling was also used to assess the relative proportions of clay minerals present in the <2 µm fractions by comparison of sample XRD traces with Newmod-for-Window™ modelled profiles. The modelling process requires the input of diffractometer, scan parameters and a quartz intensity factor (instrumental conditions), and the selection of different sheet compositions and chemistries. In addition, an estimate of the crystallite size distribution of the species may be determined by comparing peak profiles of calculated diffraction profiles with experimental data. By modelling the individual clay mineral species in this way, mineral reference intensities were established and used for quantitative standardization following the method outlined in Moore & Reynolds (1997).
Cosmogenic Isotope Analysis

In terms of determining denudation rates, $^{10}$Be has been the most widely used nuclide due to its radioactive nature; its abundance in silicate rocks and sediments; its resistance to chemical weathering; and its simple target chemistry (Gosse and Phillips 2001; von Blanckenburg, 2005). It is these factors that often make it the mineral of choice as they result in uniform cosmogenic nuclide rates and for this reason cosmogenic $^{10}$Be measurements were used in this project to determine long-term erosion rates for a high-relief mountain catchment. $^{10}$Be accumulates at a rate directly proportional to the length of time for which a surface has been exposed to cosmic rays, and inversely proportional to the rate of erosion (Lal 1991). The roughly exponential decrease in $^{10}$Be production rates with depth essentially means that the accumulated equilibrium $^{10}$Be concentration in a mineral is a function of the rate at which that grain has been uncovered, i.e. the erosion rate (Lal 1991). In-situ erosion rates may be calculated from $^{10}$Be concentrations in bedrock, and mean catchment erosion rates calculated from the $^{10}$Be content of detrital sediment provided that erosion is steady, and that catchment-wide $^{10}$Be production is balanced by $^{10}$Be leaving the catchment (i.e. there is no significant sediment storage) and that the quartz is uniformly distributed within the source area (Lal 1991; Brown et al. 1995; Bierman and Steig 1996; Granger et al. 1996; Stone 2000). Practical limits on sample size and the removal of chemical impurities constrain calculable erosion rates to $<10\text{mm.a}^{-1}$ (Reinhardt, 2007).

The duration of exposure of bedrock at the Earth’s surface may be calculated provided that the $^{10}$Be content if the surface was zero at initial exposure and that subsequent erosion was negligible (Lal 1991). The cosmogenic nuclide production rates for any site on the Earth’s surface can be calculated, taking production rate corrections such as latitude, altitude and the geometry of surrounding obstructions into account (Dunne et al. 1999; Stone 2000).

The application of cosmogenic erosion rates

This study utilized $^{10}$Be concentrations in bedrock and fluvial sediment samples to estimate surface erosion rates, along with lacustrine deposits of quartz from the two shallow water cores (MG3 & MG4) to estimate palaeo-erosion rates for the Lake Mugecuo catchment during the last 3,400 years. The following section outlines the methods used to measure $^{10}$Be concentrations and their subsequent conversion to erosion rates. The application of cosmogenic erosion rates in a depositional lacustrine setting is still relatively novel (see Schaller et al., 2002). Cosmogenic $^{10}$Be measurements in river-borne quartz sand record a time-integrated erosion rate representative of an entire drainage basin and when sequestered in a deposit of known age, palaeo-erosion rates may be recovered from the nuclide concentration of the sediment (Bierman & Steig, 1996; Granger et al., 1996; Schaller et al., 2002). Thus, this technique can be used to directly measure $^{10}$Be concentrations that are representative of mean catchment wide rates at the time of burial, allowing for post-deposition radionuclide production and decay. This technique does not require knowledge of sediment deposition rate nor depositional environment and thus provides an independent measure that can be compared with other more standard palaeoenvironmental proxies.

A prerequisite for the application of in-situ nuclides for the study of erosional histories of surfaces is a knowledge of their production rates under different irradiation conditions including; altitude, latitude, irradiation geometry and shielding (Lal, 1991), but the full quantification of the denudation rates of an entire landscape would require the analysis of a large number of representative soil and bedrock samples from various geomorphic settings in that landscape (e.g. hillslopes, river valleys, gorges, landslides and bedrock
surfaces) (von Blanckenburg, 2005). However, the measurement of cosmogenic radionuclides (CRNs) in fluvial and alluvial sediments has been shown to yield spatially averaged erosion rates and has become the method with the greatest applicability in measuring erosion rates over $10^3$-$10^5$ yr and across a wide variety of landscape and erosional processes (Granger & Smith, 2000, Bierman & Steig, 1996). Basin-averaged denudation rates are estimated from the concentrated in-situ produced cosmogenic nuclides in rocks and sediments, collected within a stream channel, or from recent deposits. Crucially, this sediment is assumed to be a fully mixed amalgam of grains that are released from all parts of the landscape in proportion to their long-term denudation rate (Norton, 2009). The nuclide concentration measured from a river sediment sample is then a mean of the contributed concentrations from all points in the watershed; at any given point the nuclide concentration will depend on the production rate and denudation rate at that location. The approach of ‘let nature do the averaging’ leads to a denudation rate that is averaged over all geomorphic processes contained in the drainage basin. Nuclides are produced in a catchment with the area $A$ "In"; and they are exported by river sediment “Out” (see Figure 2.16). If the inbound flux by production equals the outbound flux by denudation the basin is at cosmogenic steady state and the mass flux $dM/dt$ [tons year $^{-1}$] can be calculated. Dividing the mass flux by the catchment area and the rocks density results in the catchment-wide denudation rate $e$ [mm ky$^{-1}$].

![Illustration of the 'let nature do the averaging' principles of the catchment approach (von Blanckenburg, 2005).](image)

Numerical modelling of present day $^{10}$Be production allows concentrations in bedrock and fluvial samples to be interpreted either as a 'steady erosion rate or exposure age (Lal, 1991). Practical limits on sample size and the removal of chemical impurities constrain calculable erosion rates to $<10$mm.a$^{-1}$ but provided that the enough grains are sampled, mean erosion rates may be estimated in any landscape experiencing regular (shallow) landsliding over a range of depths not exceeding 3m, or $<60$ cm in width (Reinhardt et al., 2007b).
2.6.2 Fluvial and palaeo-deposit sand collection

The $^{10}$Be concentrations of fluvial quartz are used here to calculate mean detrital erosion rates following Bierman & Steig (1996) and Granger et al. (1996). Two modern fluvial sediment samples were collected at Lake Mugecuo in July 2010 (for sample descriptions see Appendix A.1.2) to determine the average modern erosion rate in the catchment. At the fluvial sediment sites (see Figures 2.17-2.18), ~10 kg of 0.25-0.5 mm alluvium sediment was collected from the surface layer of the small delta below the river outlets. The samples were sieved in the field to collect 0.25-0.5mm fractions.

Figure 2.17 Sampling sites of bedrock, fluvial, and palaeo-deposit material for $^{10}$Be concentration measurement in the Lake Mugecuo catchment. The lake bathymetry is included for reference.
Figure 2.18 Panels 1 and 2 represent fluvial sampling site B. Panel 3 shows the sieving of alluvium samples in the field to collect 0.25-0.5mm fractions. Panel 4 highlights the location of fluvial sampling site A.
In addition, 5 sand samples were taken from the two shallow water cores in order to estimate mean catchment-wide palaeo-erosion rates from the cosmogenic $^{10}$Be concentration palaeo-deposits of quartz-rich sands. When sequestered in a terrace of known age, palaeo-erosion rates may be recovered from the nuclide content of the terrace material (Schaller et al., 2002). Such erosion rate estimates are not sensitive to changes in the depositional environment. Thus, knowledge of the $^{14}$C radiocarbon dates of each of the palaeo-deposits of sand were used to develop a chronology for the pattern of erosion in this catchment during the Late Holocene. The sand-rich shallow-water lake sediment cores extracted are perfect for the type of $^{10}$Be analysis proposed here as they provide evidence of channel avulsions in the catchment. An avulsion is the natural process by which flow diverts out of an established river channel into a new permanent course on the adjacent floodplain (Slingerland & Smith, 2004) i.e. it is a natural abrupt relocation of a river. These natural relocations of a river channel provide a means of assessing changes in palaeo-denudation rates in the catchment over time as the bands of sand found in the shallow water cores likely represent a change in the fluvial system over time which may have been caused by major discharge diversions. The sand layers found in the shallow water cores (see section 3.1.2, Figure 2.19) were therefore portioned into units to extract $^{10}$Be palaeo-erosion rates based on careful consideration using particle size data and the minimum amount of pure quartz required from each sample (>30g pure clean quartz). For full sample descriptions see appendix A.1.2.

![Figure 2.19](image-url) The sand layers found in the shallow water cores which were carefully selected and portioned into units to extract $^{10}$Be palaeo-erosion rates.
2.6.3 Bedrock sample collection

Samples for $^{10}$Be analyses were collected from in-situ bedrock exposures at Lake Mugecuo. Physical erosion occurs by the detachment of particles ranging in size, thus the $^{10}$Be concentration of a surface experiencing regular detachment of a grain or slab of fixed thickness and time interval varies between being greater or less than the actual mean value (Small et al., 1997; Reinhardt et al., 2007). However, if a sufficient number of bedrock samples are collected from rock surfaces at different stages of the weathering cycle, the full range of nuclide concentrations is sampled, and the actual erosion rate may be estimated by amalgamating these bedrock samples (Reinhardt et al., 2007). 10 samples of vein quartz were therefore collected (all at the same altitude) from a single hill slope exposure on the north eastern side of the lake catchment (Figure 2.20), enabling the calculation of mean hill slope bedrock exposure.

![Image 1](image1.jpg)

**Figure 2.20** In-situ bedrock exposure sampling site at Lake Mugecuo. Image 1 provides a close-up view of the exposure. The second image is taken from the southern banks of the lake looking toward the exposure site (marked with an X) on the north eastern side of the lake catchment. The third image provides a close-up view where a vein quartz sample was collected from the hillslope.

2.6.4 Sample preparation

The following protocol was used to extract pure quartz from each of the samples outlined above and to prepare $^{10}$Be targets for AMS measurement. Protocols for this procedure vary widely depending on study sites and the lithologies investigated, therefore it was necessary to develop an independent protocol for the Mugecuo catchment. The following protocol was adapted from a variety of sources (e.g. Bierman et al., 2002; Bookhagen, 2009).

Each of the bedrock samples were cut into equal rectangular blocks with the z-axis perpendicular to the exposure face. Each sub-sample was then amalgamated, crushed and sieved into <0.5 and <0.25mm fractions. The fine alluvium sediment samples and sand-rich shallow-water lake sediment units were wet
sieved in the laboratory to 0.125-0.5mm and dried overnight at 70°C. A variety of fractions were bagged and weighed, but only the 0.25-0.5mm grains from each of the samples were used in the AMS $^{10}$Be measurement as this is the most commonly used fraction for erosion rate measurement (c.f. Bierman et al 2002). A small amount of 1% H$_2$O$_2$ was used to test for the presence of organic material (a vigorous reaction is observed if organics are present), and where significant, samples were treated with 10% H$_2$O$_2$ for 24hrs on a hotplate at 100°C.

Heavy liquids were used to separate minerals according to their density. Sodium metatungstate solution (SMT) can achieve an operating density up to 2.90g ml$^{-1}$ at 25°C and can separate quartz from mafics and other minerals with a greater density. The samples were mixed with a SMT solution which had been reduced down to a density of ~2.6 g ml$^{-1}$ (the approximate density of quartz) and passed through a separating funnel to float off any mafic materials present in the samples. This resulted in quartz particles sinking to the bottom of the separating funnel which were then released from the funnel into a clean beaker while lighter mafic material was floated off from the top of the solution.

Each sample was then etched in a hydrochloric (HCl) and hydrofluorosilicic ($\text{H}_2\text{SiF}_6$) acid mixture to rid the samples of carbonate material and iron coatings on grains as well as all other minerals other than silicates i.e. quartz. 100ml HCl was added to the samples in labelled 1l nalgene bottles. As long as no fizzing was observed, 200ml of $\text{H}_2\text{SiF}_6$ was added to the sample which was then placed in an ultrasonic bath and agitated overnight at 50°C.

As per Kohl and Nishiizumi (1992), weak HF/HNO$_3$ (Nitric) acid was then used to progressively leach out non-quartz while also removing the outer shell of each quartz grain thereby stripping atmospheric beryllium from the surface of the grains. For this purpose each quartz sample was subdivided into ~100g aliquots and placed into 1l nalgene bottles. These aliquots were then etched in 90ml 40% HF/ 50ml 70% HNO$_3$ for 24hrs on a shaker table.

It was found that the quartz samples from the Mugecuo catchment contained significant quantities of feldspars which cause the samples to have extremely high aluminium levels and are essentially artefacts that are left in the samples that the acid had failed to remove. In some cases minute quartz veins are embedded in feldspar substrate, or appear as a thin surface veneer, which makes extraction and processing of the fragile quartz component difficult (Mifsud et al., 2013). Testing for aluminium concentrations in the samples is essentially used as a purity test as high concentrations interfere with the $^{10}$Be analysis. Mild crushing (as per Shen et al. 2009) i.e. less than one minute using a pestle and mortar, was necessary to allow the acid to leach out these feldspars, thus lowering the aluminium concentrations in the samples to <100ppm.

In order to determine the aluminium concentration of each of the quartz samples, a 1g sub-sample was placed into a 7ml Savillex beaker. Each sample was then dissolved in 3ml conc. HF overnight, dried down on a hotplate (~90°C), and then re-dissolved in 1ml conc. HNO$_3$. After heating for 1hr, the solution was left to evaporate. If any material remained undissolved, aqua regia (2ml 6M HCL and 1ml 15M HNO$_3$) was added to the sample and evaporated. 1ml 0.3 M HNO$_3$ was then added to each beaker. When all of the sample had dissolved 1ml of the remaining solution was decanted into a centrifuge tube and 9ml 0.3M HNO$_3$ was added and shaken. One or more blank solutions were prepared in order to act as a control by decanting 10ml of the
0.3 M HNO₃ solution into empty tubes to test the consistency of the laboratory results. These solutions were then commercially measured using ICP-AES analysis. The aluminium ppm concentrations were calculated using the following formula:

\[
\text{Alppm} = \text{ICP result} - \text{Blank} \times \frac{\text{HNO}_3 \text{ weight}}{\text{quartz weight}}
\]

If more than 100ppm of aluminium was present, the leaching HF/HNO₃ acid stage of the protocol was repeated until acceptable Al concentrations were achieved in order to proceed with the Be extraction and purification procedure.

Due to limitations in laboratory facilities, the final steps in the protocol were carried out by Dr. David Fink at the Australian Nuclear Science and Technology Organisation (ANSTO) in Sydney, Australia following the protocol outlined in Mifsud et al. (2013). The natural concentrations of \(^{10}\text{Be}\) are too low to be detected by AMS, therefore a known concentration of \(^{9}\text{Be}\) was added to each of the ‘clean’ quartz samples and dissolved in HF. Anion exchange chromatography and cation exchange chromatography were subsequently used to separate Be from the pure quartz samples. Be hydroxide was then precipitated from the separate and transformed into BeO to be measured by accelerator mass spectrometer at the AMS facility at the Australian Nuclear Science and Technology Organisation (ANSTO) in Sydney, Australia by Dr. David Fink.

### 2.6.5 Production rate scaling and topographic shielding

The calculation of erosion rates using cosmogenic isotope measurements depends on precise knowledge of local isotope production rates which are calibrated from a small number of sites around the world. These calibration sites are of known surface exposure age and must be scaled to the sample site by using factors that correct for differences in cosmic ray shielding by the Earth’s magnetic field and altitude-dependent shielding by the atmosphere (Stone, 2000). For the procedure of calculating on site \(^{10}\text{Be}\) production rates, Stone (2000) gives scaling factors for latitude and altitude based on those of Lal (1991), with two changes: (1) they are recast in terms of atmospheric pressure rather than altitude (i.e. production rates are dependent upon atmospheric depth rather than altitude), and (2) by assuming a 3% contribution of muons to the production of \(^{10}\text{Be}\) at sea level (instead of the >15% used by Lal, 1991), the scaled production rates from the different sites worldwide to converge to approximately 4.49 ± 0.39 \(^{10}\text{Be}\) atoms.g\(^{-1}.a\(^{-1}\), thereby solving the discrepancy between published production rates (Gosse and Phillips, 2001). International projects such as CRONUS and CRONUS-EU (Dunai, 2005) are working towards increased precision in cosmogenic nuclide methods and analysis techniques; the goal of these initiatives is to reduce errors associated with production rates, scaling and analysis to 5 per cent. Currently, these errors are between 10 and 20 per cent, but as analytical and theoretical errors continue to be reduced, the relative importance of the common correction factors, such as topographic and snow shielding increases; therefore, any method for calculating these factors should aim to be more accurate than the 5 per cent goal of these studies (Norton, 2009).

Topography contributes to the reduction of \(^{10}\text{Be}\) production by (1) shielding a certain proportion of the incoming radiations and (2) modifying the effective attenuation length by changing the angle of incidence on sloping surfaces (Dunne, 1999, Gosse and Phillips, 2001). Topographic shielding becomes an important issue in locations where the land surface is rough (i.e. in high relief areas such as the Lake Mugecuo Sichuan field site) and an accurate assessment of topographic shielding would include measurement of the
angle to the horizon in all directions from every point in the basin (Norton, 2009). Scaling factor is defined as the ratio of the shielded to the total (or maximum) cosmic ray flux which would be produced in a flat unshielded catchment (i.e. the flux received by a horizontal, unobstructed surface) (Codilean, 2006). Any matrix-based system can be used to determine the inclination to the horizon in all directions from every pixel in a watershed using digital elevation models. From this, a representative topographic shielding factor can be calculated, and the production rate scaled accordingly. However, DEM grid size has a marked effect on the topographic shielding factor calculated for a surface. The magnitude of the effect of grid size on topographic shielding factors varies depending on the roughness of the terrain; rougher terrains provide higher topographic shielding, and the rate of change with increasing cell size is larger for highly variable terrains than for flatter terrains (Norton, 2009). This is an important factor that was taken into account when calculating production rates for the field site at Lake Mugecuo which stands at an elevation of approximately 3,700m. In this study a 90m resolution digital elevation model DEM was used to calculate production rates for the altitude of every cell following Granger et al. (2000, 2001), Stone (2000) and Balco (2008) and corrected for topographic shielding (Codilean, 2006). The relationship between the topographic shielding factor and grid size is a log linear function of the form:

$$S_{basin} = a \ln(G) + S_0$$

where $S_{basin}$ is the basin-averaged topographic shielding factor, $G$ is the grid size (m), $a$ is a coefficient related to the local variability in surface elevation and $S_0$ is the theoretical maximum shielding (Norton, 2009). The rate of decrease of topographic shielding with increasing grid size is a function of topographic roughness. All measurements of topographic roughness are dependent to some degree on the distance or scale over which roughness is measured but this dependence can be minimized by using the entire analysis area (i.e. the drainage basin) as the measurement scale (Zhang, 1999). The 90m resolution raster DEM of the Mugecuo catchment was used to calculate production-rate calculations and the following variables for every cell in the DEM based on the methods described in Balco et al. (2009):

1) Slope
2) Spallation production scaling factor (Stone, 2000)
3) Muon scaling factor (Stone, 2000)
4) Combined scaling factor (Stone, 2000)
5) Topographic shielding factor (Dunne 1999) (bedrock samples only)

The correction for sample thickness was calculated as per Balco (2001) and the correction for topographic shielding at the outcrop site at Lake Mugecuo site was calculated as per Dunne (1999), using the dip angle of the horizon over each homogeneous interval of the azimuth, and for surface inclination.

2.6.6 Conversion of $^{10}$Be rates to palaeo-erosion rates

A wide range of assumptions are required for the calculation of steady erosion rates from $^{10}$Be concentrations in quartz as outlined in Chapter 1. Numerical modelling (see Appendix A.1.3 for calculations) of present day and palaeo $^{10}$Be concentrations allows concentrations in bedrock and fluvial quartz to be
interpreted in terms of a steady erosion rate (Lal, 1991; Granger et al., 1996; Balco et al., 2008). From the measured $^{10}\text{Be}/^9\text{Be}$ ratios, blank corrected $^{10}\text{Be}$ concentrations in quartz are calculated based on the weight of quartz dissolved and the amounts of carrier added ($^9\text{Be}$). These concentrations (with fully propagated errors) are then used for erosion-rate calculations. Propagated errors ($1\sigma$) include the analytical error based on counting statistics and the error of the blank correction. The methods and terms described in Balco et al. (2009) were followed for the steady erosion-rate calculation methods incorporated in the CRONUS-Earth online exposure age calculator, Version 2.2 (Balco et al., 2008). See Table 2.1 for input parameters for the calculator. Note that the version 2.2 of the Cronus-Earth calculator (Balco et al., 2008) uses a $^{10}\text{Be}$ half-life of 1.39 ± 0.1 × 10$^6$ yr (Chmeleff et al., 2010; Korshinek et al., 2010). The reference $^{10}\text{Be}$ production rate for spallation is 4.49 ± 0.39 atoms g$^{-1}$ yr$^{-1}$ for this scaling scheme and the muonic production rate is based on measurements by Heisinger (2002a, b; see also Balco et al., 2008, for further details). The total measurement error in $^{10}\text{Be}$ concentration includes the 1 σ AMS uncertainty measurement as well as additional 1 σ error associated with sample preparation and analysis (c.f. Gosse and Phillips 2001).

Table 2.1 Input parameters for CRONUS-Earth online exposure age calculator, Version 2.2 (Balco et al., 2008)

<table>
<thead>
<tr>
<th>Sample name</th>
<th>Latitude (DD)</th>
<th>Longitude (DD)</th>
<th>Atmospheric Pressure or Elevation</th>
<th>Elv/pressure flag</th>
<th>Thickness (cm)</th>
<th>Density (g cm$^{-2}$)</th>
<th>Shielding correction</th>
<th>Be standardization</th>
</tr>
</thead>
<tbody>
<tr>
<td>MG4a</td>
<td>30.157</td>
<td>101.5</td>
<td>600.21</td>
<td>pre</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>NIST_Certified</td>
</tr>
<tr>
<td>MG4b</td>
<td>30.157</td>
<td>101.5</td>
<td>600.21</td>
<td>pre</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>NIST_Certified</td>
</tr>
<tr>
<td>MG3a</td>
<td>30.156</td>
<td>101.5</td>
<td>600.21</td>
<td>pre</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>NIST_Certified</td>
</tr>
<tr>
<td>MG3b</td>
<td>30.156</td>
<td>101.5</td>
<td>600.21</td>
<td>pre</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>NIST_Certified</td>
</tr>
<tr>
<td>MG3c</td>
<td>30.156</td>
<td>101.5</td>
<td>600.21</td>
<td>pre</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>NIST_Certified</td>
</tr>
<tr>
<td>MGDR</td>
<td>30.159</td>
<td>101.54</td>
<td>600.21</td>
<td>pre</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>NIST_Certified</td>
</tr>
<tr>
<td>MGMR</td>
<td>30.155</td>
<td>101.53</td>
<td>632.38</td>
<td>pre</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>NIST_Certified</td>
</tr>
<tr>
<td>MGCR</td>
<td>30.155</td>
<td>101.55</td>
<td>3831</td>
<td>std</td>
<td>0.960453</td>
<td>2.6</td>
<td>0.45</td>
<td>NIST_Certified</td>
</tr>
</tbody>
</table>

Note: Elv/pressure flag specifies how to treat the elevation/pressure value. When a standard atmosphere is applicable at the site (locations outside Antarctica), the text string "std" is used. When using pressure in hPa, the text string "pre" is used. NIST_Certified refers to a standard produced by the National Institute of Standards and Technology (NIST), referred to as SRM4325, with the nominal isotope ratio stated on the certificate for this material (2.68 × 10^{-4} for the solution as supplied by NIST).

The following formula (Lal, 1991; Small et al., 1997) was also used to cross-check the calculations of steady erosion rates from $^{10}\text{Be}$ concentrations using the CRONUS-Earth online exposure age calculator, Version 2.2 (Balco et al., 2008). Present day and palaeo-erosion rates were calculated using the formula below in conjunction with the scaling factors calculated using numerical modelling in Matlab v. R2011 (see Appendix A.1.3 for script used).

$$ E = \left( \frac{P(0)}{N_k} - \lambda k \right) \frac{\Lambda}{\rho} $$

Equation 1. The calculation of 10Be erosion rates (Lal, 1991; Small et al., 1997)

Where E is the erosion rate (cm a$^{-1}$), P(0) is the average TCN surface production rate, Nk is Avogadros number (6.022 x 1023 atoms mol$^{-1}$), $\rho$ is the mean rock density (2.6 ± 0.05 g cm$^{-3}$; Zhang et al. 2002), $\lambda k$ is the cosmogenic isotope decay constant and $\Lambda$ is the latitude dependent mean free path length for spallation.
(g cm⁻³; taken from Lal, 1991, Balco, 2008). The scaling scheme of Lal (1991) was used here along with Stone (2000) for the spallation production rate.

¹⁰Be analysis works very well on young or well-preserved terraces. Difficulties arise on degraded terraces where soil formation, bioturbation, and sediment transport have disturbed the surface (Granger & Smith, 2000). The palaeo-erosion rate samples in this study are relatively young and well-preserved therefore no allowances for post-deposition radio-nuclide production and decay had to be made. The shielding effect of the palaeo-samples by water was found to be negligible (c.f. Bierman & Nichols, 2004) and is therefore not considered here.

2.7 Pollen Analysis

Von Post (1916) first identified that there is a relationship between pollen archived in a sedimentary profile and the vegetation occurring in the surrounding area. Palaeoecology is defined by Rull (2010) as 'the branch of ecology that studies past ecological systems and their trends in time using fossils and other proxies'. Studies of ecological systems, past, present, or future, require consideration not only of their biotic components (e.g. species, populations, communities, biotic interactions, human activities) but also their abiotic components (e.g. climate, soil, topography, water chemistry, water temperature) (Birks & Birks, 2006).

Pollen grains are the most abundant and well-dispersed fossil plant remains and analysis of sedimentary pollen records is one of the most powerful methods for reconstructing past vegetation and climate (Meltsov et al., 2011). Pollen analysis is also an effective tool to detect and understand the effect of climate and land-use changes on land cover composition at various time scales (Elenga et al., 2000; Tarasov et al., 2000; Seppä et al., 2003; Nielsen and Odgaard, 2005; Marchant et al., 2009; Gaillard et al., 2010). Sedimentary records of fossil pollen and larger plant remains offer a time-series of range adjustments of plant species and potentially give insight into processes important for maintaining biodiversity at medium to long time-scales (Odgaard 2001). Pollen records reflect flora and vegetation at a variety of spatial scales (Birks, 1986), and it is important to remember that most existing pollen records reflect a mixture of communities and/or successional stages (Odgaard, 1999).

Lacustrine pollen is an important proxy for reconstructing palaeo-vegetation and palaeo-climate, because pollen grains are usually well preserved in lake sediments (Sun & Wu, 1987; Gasse et al., 1991; Chen et al., 2006). The regional pollen rain reflects vegetation at a regional scale, but pollen may also be derived more locally, such as from lake-side and aquatic vegetation (Birks & Birks, 2006). Pollen dispersal ability, dispersal modes (e.g. wind or insect pollinated), basin size, site type, plant height, wind speed, and existence of instreams all influence the pollen source area of a lake catchment (Jacobson & Bradshaw, 1981; Prentice, 1985; Sugita, 1993, 1994; Jackson, 1994; Jackson & Lyford, 1999). The size of depositional basin is the one factor suggested most to influence the size of a source area (Jacobson & Bradshaw, 1981; Prentice, 1985; Sugita, 1993; Jackson & Lyford, 1999; Davis, 2000). The larger the catchment size, the greater the proportion of pollen loading coming from regional sources; the lighter the pollen grain, the greater the distance of pollen transport in the air; and the closer the source plants, the higher the pollen deposition from these plants (Poska et al. 2011). In small lake basins, more plants are within a short distance to the deposition site, and contribute more pollen grains to the pollen spectrum (Weng 2006). In mountain environments where vegetation zones are compressed into narrow belts by increasing elevation, long-
distance transport of pollen can be problematic (Maher 1963; Birks 1977; Kearney 1983; Evans 1997; Pellatt et al. 1997). Further complications arise when pollen grains are lost due to poor preservation, differential production of species in response to habitat or climatic changes, sediment focusing and changing sedimentation rates associated with limnological and catchment processes (Davis et al., 1973; 1984). However, high elevation lakes are generally undisturbed by human activities, thereby providing ideal archives for the study of past climates (Ortu et al., 2006).

Mountainous regions are likely to be the first place to exhibit a response to future climate change (Bunting, 2008). Tinner et al. (2005) investigated the response of high mountain vegetation to environmental changes in the Swiss Alps and discovered that tree line vegetation was in dynamic equilibrium with climate, including during abrupt dynamic shifts. Other studies have shown that the response of forests to climate change in mountain environments can have a lag of up to 150 years (Davis and Botkin, 1985; Bunting, 2008). This suggests that care must be taken when interpreting a pollen record as shifts in arboreal pollen may be related to an event that occurred earlier than implied in the pollen archive. It is also important to note that certain events may not be recorded in a pollen record if they were not of sufficient magnitude to trigger a vegetational response in the catchment. Thus, a ‘smoothing’ of climate changes recorded in a pollen record can often occur, and low frequency or even high magnitude climatic event may not be recorded in the lake archive. However, a high altitude location can give rise to strong altitudinal/environmental gradients that exist over relatively short distances resulting in the catchment being highly sensitive to climatic changes (Bunting et al., 2008).

Fossil pollen data are used to infer past floral diversity based on the premise that at the landscape scale, pollen and plant diversity are roughly correlated (Moore, 1973; Seppä, 1998; Weng et al., 2007). The number of pollen types in sediment samples, the so-called palynological richness (Birks and Line, 1992) is one of the most simple and widely used proxies of past vegetation diversity (Seppä, 1998; Veski et al., 2005; Berglund et al., 2008; Saarse et al., 2009; Valsecchi et al., 2010). Pollen diversity differs from plant diversity owing to the lower taxonomic precision of pollen identification and ‘silence’ of many species, especially entomophilous taxa, which are not represented in pollen spectra (Faegri & Iversen, 1989). Species diversity is composed of two components- species richness and species evenness (Ludwig and Reynolds, 1988). Species richness is most simply represented as the number of species in the community (Laird et al., 2010), while species evenness is the distribution of species abundances (Legendre and Legendre, 1998). Indices that attempt to combine species richness and evenness into a single value are referred to as diversity indices. These diversity indices can confound important characterizations of community structure and ultimately can lead to loss of information (James and Rathburn 1981) so caution must be exercised when using such indices. The reconstruction of plant community dynamics based on fossil pollen data are further discussed in the following section.

2.7.1 Basic pollen data analysis

Pollen was extracted from freeze-dried samples from MG1 (the deep water core) at 2cm intervals following standard techniques described by Faegri and Iversen (1981) and Moore et al. (1991). Sub-sampling for pollen analysis was focused on the sediment from the channel laid down in the middle of the coring tubes as opposed to the distorted sediment either side of this channel to avoid the distortions caused by problems during core extraction. Pollen identification was undertaken by Professor Tong, Department of Hydrology.
and Environmental Geology, Chinese Academy of Geological Science (CAS), Shijiazhuang. Thus, these data are in effect secondary data from the point of view of this thesis, though the author was responsible for sampling design and collection.

Raw pollen counts of trees, shrubs, herbs, grasses, sedges, fern and aquatics were converted to percentages using Tilia v2.0.2. (Grimm, 2004) which are useful for determining the contributions of taxa to the pollen record. The pollen sum must always equal 100% which means that % estimates do not take into account the differing rates of pollen production and dispersal of each taxon (Moore et al., 1991). Hence, some taxon may be over or under-represented in the pollen record. The investigation of pollen concentrations are useful for overcoming the inherent issues associated with pollen percentages as these can provide an indication of the relative abundances of taxon over time (Smol et al., 2001a). To calculate pollen concentrations for each taxon, two tablets containing a known concentration of Lycopodium spores were added to each pollen sample prior to pollen extraction and identification. Once the number of pollen and spores had been counted in each sample, pollen concentrations could be calculated following Stockmarr (1971). Pollen frequency and concentration diagrams were produced using TGview software version 2.0.2. (Grimm, 2004).

\[
\text{Equation 2. Pollen concentration calculation (Stockmarr, 1971)}
\]
\[
\text{No. Grains Counted} \times \text{No. Lycopodium spores added} \times \frac{1}{\text{No. Lycopodium counted}} \times \frac{1}{\text{Sample Mass (g)}}
\]

Statistical analysis involving detrended correspondence analysis (DCA), principal components analysis (PCA) and ratios of one taxon to another have proven useful tools for semi-quantitative assessments of variables such as precipitation and/or temperature shifts (Zhao et al., 2009; Herzschuh, 2007; Van Campo et al., 1993). Taxa with abundances >0.5% were subjected to numerical analysis using CANOCO v. 4.5.2 (ter Braak and Smilauer, 2003). Detrended Canonical Analysis (DCA) was used to explore the main patterns of taxonomic variation among samples and to estimate the compositional gradient lengths along the first few DCA axes. The results of this analysis are expressed as axes and related gradients (Frey et al., 1998). The results of the DCA analysis pointed to a linear response of species data to environmental gradients (Birks, 1998). The dataset was subsequently analysed using Principal Components Analysis (PCA), an indirect linear analysis of the species data which summarizes variation in the record (see Appendix A.1.4 for details on PCA parameters/analysis output). The ordination of pollen taxa allows the integration of different pollen taxa into groups with similar response to the (theoretical) environmental gradients (Kramer et al., 2010). The data were square-root transformed, rare samples were down-weighted and centering/standardisation was by species. The pollen percentage species data were transformed using square roots in an attempt to stabilize the variance and to minimize the ‘signal to noise’ ratio in the data (Prentice, 1980; Zhang, 2002; Seppä et al., 2004).

2.7.2 Interpretation of changes in pollen diversity/evenness

In spite of a long history of research and progress in interpreting fossil pollen records, it is still uncertain how these data can be used as a precise record of floristic and vegetation diversity dynamics (Odgaard, 1999). It has often been hypothesised that the number of pollen types in sedimentary assemblages is directly
proportional to the floristic diversity of the pollen catchment area (MacDonald et al. 2008), but the relationship between palynological diversity and floristic diversity is still incompletely understood (Odgaard, 2001). The number of pollen types in sediment samples, the palynological richness (Birks and Line, 1992) is one of the most simple and widely used proxies of past vegetation diversity (Seppä, 1998; Veski et al., 2005; Berglund et al., 2008; Saarse et al., 2009; Valsecchi et al., 2010). Species diversity is composed of two components- species richness and species evenness (Ludwig and Reynolds, 1988). Species richness is most simply represented as the number of species in the community (Laird et al., 2010), while species evenness is the distribution of species abundances (Legendre and Legendre, 1998). Indices that attempt to combine species richness and evenness into a single value are referred to as diversity indices. These diversity indices can confound important characterizations of community structure and ultimately can lead to loss of information (James and Rathburn 1981). Diversity indices such as the Simpson and Shannon-Weiner indices have been widely used to infer past floral diversity (Cwynar, 1982; Ritchie, 1982; Cao & Zhang, 1997; Chen et al., 2006; Weng et al., 2007). However, they do not account for sample size which is where rarefaction becomes a useful tool in ensuring a standardized number of taxa for each pollen count which then provides us with an estimation of pollen richness (Laird et al., 2010). In an ideal situation, all pollen taxa are detected in a pollen investigation, and temporal variations in detected palynological richness reflect the true changes in plant diversity in the vegetation (Weng et al. 2006). Unfortunately, the detected richness in pollen spectra tends to vary with pollen count i.e. the correlation between plant and pollen is not a one-to-one relationship. To address this problem, Birks and Line (1992) introduced the concept of rarefaction, a technique that permits an estimation of pollen richness using standardized pollen counts. This standardization is hugely important as palynological richness depends in part on pollen sum- as the sum increases, more taxa can potentially be detected (Peros and Gajewski 2008). Rarefaction analysis implements such a standardization and provides minimum variance unbiased estimates of the expected number of taxa \( t \) in a random sample of \( n \) individuals taken from a larger collection of \( N \) individuals containing \( T \) taxa. The units of rarefaction analysis represent the estimated number of pollen taxa reduced to a certain pollen sum, usually the lowest sum of the group of samples being compared (Peros & Gajewski, 2008). The estimated number of taxa \( E(T_n) \) is given by:

\[
E(T_n) = \sum_{i=1}^{T} 1 - \frac{(N - N_i)! (N - n)!}{(N - N_i - n)! N!} 
\]

Equation 3. Rarefaction Analysis (Birks and Line 1992; Weng et al. 2006)

Where \( E(T_n) \) is the expected palynological richness in a standardized pollen count \( N_i \), \( T \) is the palynological richness in the original pollen count, \( N \) is the overall pollen sum, and \( n \) is the number of grains selected for standardization in the rarefied sample (Birks and Line 1992; Weng et al. 2006). It thus removes the unavoidable bias in richness estimates caused by differing count sizes. Therefore, a suitable comparison of species richness estimated from samples of different sizes can be made only after ‘rarefaction’ to the same number of plots (see Figure 2.21 taken from Gotelli and Colwell, 2001).
Figure 2.21 Sample- and individual-based rarefaction and accumulation curves. Accumulation curves (jagged curves) represent a single ordering of individuals (solid-line, jagged curve) or samples (open-line, jagged curve), as they are successively pooled. Rarefaction curves (smooth curves) represent the means of repeated re-sampling of all pooled individuals (solid-line, smooth curve) or all pooled samples (open-line, smooth curve). The smoothed rarefaction curves thus represent the statistical expectation for the corresponding accumulation curves. The sample-based curves lie below the individual-based curves because of the spatial aggregation of species. The individual-based accumulation curve shows one particular random ordering of all individuals pooled. The sample-based accumulation curve shows one particular random ordering of all samples in the dataset. Taken from Gotelli and Colwell (2001).

Across-sample rarefaction was used to estimate the species richness of every 2cm sample in the MG1 pollen dataset for a standard number of individuals; by standardising a constant number of individuals it is possible to compare the samples for a standardised effort and thus determine if species richness varies between samples (Seaby & Henderson, 2007). The Shannon-Weiner (S-W) index was used to measure pollen diversity in the each of the 2cm sediment slices. This index accounts for both abundance and representation of the species present and characterizes their diversity within a community (Ledru et al., 2009). After the diversity index method was run, a standard bootstrap method for estimating upper and lower 95% confidence intervals was undertaken. Jack-knifing was then used to estimate the standard error associated with the estimates of the diversity index. SHE analysis was used to examine the relationship between S (Species Richness), H (S-W diversity index) and E (Evenness as measured using Pielou J evenness index (Pielou, 1975)). SHE analysis looks at the contribution of species number and equitability to changes in diversity (Seaby & Henderson, 2007). Species Diversity Richness (SDR) IV software was used to carry out all the above analyses.

2.7.3 Climatic indicators

Qualitative palaeoclimatic interpretation of the fossil pollen spectra can be extended by using indicator species and pollen ratios (Herzschuh et al., 2004). Hall and Grinnell (1919) were among the first to use the indicator concept by associating plant and animal species to particular ‘life zones’ (i.e. large geographic areas with similar structural and compositional characteristics) (Carignan & Villard, 2002). Previous studies have used selected arboreal taxon as climatic indicators (Herzschuh et al., 2004; Shen et al., 2005). In this study evergreen Quercus is used to infer cold, dry winters (as per Shen et al., 2005) whilst ratios were viewed as another semi-quantitative indicator of changes in vegetation types in the catchment to supplement...
the assessment of pollen analysis in the Lake Mugecuo record. The arboreal/non-arboreal pollen ratio (or arboreal pollen sum, AP) is widely used to infer changes in the landscape openness of forest steppe and sub-alpine environments and as an indicator of moisture and temperature variability in the semi-arid and mountainous areas of northern and western China (e.g. Yan et al., 1999; Liu et al., 2002; Shen et al., 2005; Herzschuh et al., 2006b). An AP/NAP ratio of 8.5 reflects dense forest, open woodland/grassland have a ratio of 4, and grassland vegetation has a ratio of ~2.5 (Herzschuh, 2007).

The *Artemisia*/Cyperaceae (A/Cy) ratio was introduced by Herzschuh et al., 2006a) as a semi-quantitative measure for summer temperature changes on the central and eastern Tibetan Plateau. It is based on the observation that cold high-alpine environments (both alpine meadows and alpine steppes) on the eastern and central Tibetan Plateau are dominated by Cyperaceae (mostly *Kobresia*), while warmer areas at lower elevations on the southern and north-eastern Tibetan Plateau are covered by temperate steppe vegetation with higher *Artemisia* coverage in the vegetation. Ratios of <1 correspond to tundra vegetation, ratios of >1 to <16 reflect steppe, and up to 26 represents steppe desert vegetation (Yu et al., 2000a). The A/Cy ratio is used here as a semi-quantitative measure for summer temperature changes in the catchment throughout the Holocene. The A/Cy ratio is the ratio of *Artemisia* to Cyperaceae pollen percentages following the method outlined in Herzschuh et al. (2006a).

### 2.7.4 Plant functional types

Predicting the effects of climate on plant community composition is a major challenge for ecologists. Plant functional type (PFT) classification has recently received new attention as a possible framework for predicting ecosystem response to human-induced changes at a global scale (Pillar, 1999). The concept of PFTs is not a new one (See Table 2.2) (Raunkiaer, 1934; Grime, 1977; Noble & Slatyer, 1980; Box, 1981). The explorations of Alexander von Humboldt (1807) in the late 1700s in Europe, Africa and the Americas led to his recognition of at least 16 species-based structural classes having different physiognomies or plant growth forms, which were not dependent on taxonomic affiliation (von Humboldt, 1849). The last decade has seen a marked increase in the availability of information on plant traits and their relationship to climate (Wright et al., 2004). The study of vegetation response to environmental change on a global scale cannot rely on species data alone as most plant species have geographically limited distributions (Pillar, 1999; Woodward & Cramer, 1996). The PFT concept consists of aggregating multiple species traits, allowing for the reduction of thousands of species to a small set of functional groups (typically <15) defined by their phenology type, physiognomy, photosynthetic pathway, and climate zone (Poulter et al., 2011). Implicit in the definition of PFTs is the idea that functional types provide a useful means of predicting the assemblage of species’ responses within an ecosystem to environmental conditions or disturbance (Hooper & Vitousek, 1997).
Table 2.2 A summary of the development of the plant functional type (PFT) concept (Duckworth 2000)

<table>
<thead>
<tr>
<th>Author</th>
<th>Concept</th>
</tr>
</thead>
<tbody>
<tr>
<td>von Humboldt (1806)</td>
<td>First recognized relationship between plant form and function. Developed classification based on growth form.</td>
</tr>
<tr>
<td>Grisebach (1872)</td>
<td>Classification of 60 vegetative forms correlated with climate</td>
</tr>
<tr>
<td>Warming (1884; 1909)</td>
<td>Classification based on simple life history features (e.g., lifespan and vegetative expansion power)</td>
</tr>
<tr>
<td>Schimper (1903)</td>
<td>Recognized convergence between plant form and function, despite taxonomic differences, between vegetation types from geographically different, but climatically similar, areas</td>
</tr>
<tr>
<td>Raunkiaer (1907; 1934)</td>
<td>Life-forms system</td>
</tr>
<tr>
<td>Kearney and Shantz (1912)</td>
<td>Proposed four basic strategies of plants in arid regions in response to drought</td>
</tr>
<tr>
<td>Braun-Blanquet (1928)</td>
<td>Added further detail to the life-forms system</td>
</tr>
<tr>
<td>Gimingham (1951)</td>
<td>Growth-forms system which also considered branching of stems</td>
</tr>
<tr>
<td>Dansereau (1951)</td>
<td>Classification system based on life-form, morphology, deciduousness and cover</td>
</tr>
<tr>
<td>Küchler (1967)</td>
<td>Hierarchical classification, with initial division based on whether plant is woody or herbaceous. Lower-order groups are based on life-forms, leaf characteristics and cover</td>
</tr>
<tr>
<td>Mooney and Dunn (1970)</td>
<td>Investigation of form–environment relationships in the context of convergent evolution</td>
</tr>
<tr>
<td>Mooney (1974)</td>
<td></td>
</tr>
<tr>
<td>Hallé et al. (1978)</td>
<td>Models of tree architecture based on the underlying ‘blueprint’ for development rather than morphology at any given moment</td>
</tr>
<tr>
<td>Box (1981)</td>
<td>Developed global classification based on structural and phenological attributes in relation to climate</td>
</tr>
<tr>
<td>Grime (1974; 1979a; 1979b)</td>
<td>Plant strategy theory and C-S-R system of PFTs</td>
</tr>
</tbody>
</table>

The concept of functional diversity offers a simplifying paradigm whereby species can be grouped together based on their physical impact on a landscape (Reinhardt et al., 2010). Whilst the importance of an existing vegetation cover in controlling erosion is widely accepted (Chirino et al., 2006; Bochet et al. 2006; Bautista et al. 2007; Durán Zuazo & Pleguezuelo 2008), little is known about the effects of plant diversity on the protecting vegetation cover (Körner 2004). The concept of PFTs can be applied at a range of spatial scales, from the plant community through the ecosystem and landscape, to regional and global scales (Ni, 2003) i.e. the major advantage of the PFT classification system is that it allows for the possibility for posing testable hypotheses that are feasible at global and centennial scales (Smith et al., 1997). What we do not yet know is what level of biological variation matters i.e. how many different types of species or functional groups play an important role in modulating and responding to landscape evolution across different temporal and spatial scales? Cardinale (2006) summarized the results of 55 experiments that manipulated the species diversity of...
plants, animals, bacteria or fungi, in a controlled setting and showed that the biomass (the standing stock abundance of all organisms in a community) tends to increase with the number of species in the community, and that the average of all curve fits suggests that the most diverse communities achieve twice the abundance or biomass of an average monoculture. Thus, to the extent that biomass impacts physical processes that shape a landscape, and the diversity of life regulates the amount and stability of biomass, then it follows that the functional diversity represented by different species plays an important role in modulating and responding to landscape evolution across scales (Reinhardt et al. 2010).

A recurring theme of ecosystem ecology is that a generalization of organisms into functional groups is needed to simplify ecosystem processes so that they can be understood and tested, in part to enable the prediction of environmental changes (Chapin et al., 1996, 1997; Smith et al., 1997). The validity of plant functional classifications has been tested in the literature using experimental (e.g., Bret-Harte et al., 2008; Keith et al., 2008), empirical (e.g., McIntyre & Lavorel, 2001; Pausas et al., 2004), and theoretical approaches (e.g., Bradstock et al., 1998; Bond et al., 2005). However, these studies do not account for vegetation dynamics (Boulangeat et al., 2012). There has been a widely recognized need to move away from single-leaf to whole-plant approaches in vegetation modelling (Körner, 1993). Models to predict vegetation responses to climate change at regional to global scales now rely on PFTs (Harrison et al., 2010). Significant efforts have been put into grouping species by functional characteristics to predict grassland (e.g., Lavorel et al., 1998) and forest ecosystem (e.g., Verheyen et al., 2003) responses to global changes. A PFT legend accompanies the MODIS land cover product using categories defined by Running et al. (1995) and has been developed from GLC2000 (Wang et al., 2006). For these particular PFT legends, the classifications include phenology type but not the associated climate zone, which is needed to assign climate-specific physiological parameters to each PFT (i.e., Sitch et al., 2003). Conversely, dynamic global vegetation models (DGVMs) (Prentice et al., 1992; Neilson, 1995) use climate variables to constrain the geographic distribution of plant functional types (PFTs) (Box, 1981, 1996; Woodward, 1987). In these approaches, biomes are only implicitly delineated from climatic constraints acting upon PFTs (Woodward, 1987; Woodward & Kelly, 1997). These groups, or functional types, are used to simplify biological diversity in dynamic global vegetation models (DGVMs) (Purves & Pacala 2008; Ostle et al. 2009) and make predictions about community and ecosystem processes (Duckworth et al., 2000). DGVMs have been used to project impacts of climate change on vegetation and carbon cycling (Cramer et al. 2001; Friedlingstein et al. 2006; Fischlin et al. 2007) but such predictions have shown considerable divergence (Sitch et al. 2008). One source of this divergence may be the models' still rudimentary treatment of PFTs. Current DGVMs recognize only about five to 15 PFTs, defined ad hoc using little of the knowledge generated during the past decade of comparative plant ecology (Harrison et al., 2010). This indicates that an enhanced PFT scheme is a major priority for improving DGVMs (Prentice et al. 2007) and other vegetation models alike.

The use of PFTs helps to solve the problem of classifying palaeoecological records a) by reducing the number of entities considered and b) by providing an ecological rather than phylogenetic basis for treating plants from different regions in a compatible way (Prentice & Webb, 1998). Classifications of PFTs can be made a priori (tentatively) or a posteriori on the basis of measured traits (Smith et al. 1997; Díaz and Cabido 2001; Violle et al. 2007). The basis of plant functional classifications and the selection of traits is dependent on the aims and context of the study and this has led to functional classifications to predict responses to
climatic changes (Woodward and Cramer 1996; Smith et al. 1997; Diaz and Cabido 1997; Box and Fujiwara 2005), small-scale disturbances (Lavorel et al. 1997; Grime 2001), or landscape-level disturbances (Noble and Gitay 1996). Functional groups have been described as groups of species that share adaptive traits for a specific function and are an important ecological framework for describing the mechanisms underlying vegetation responses (Lavorel et al., 1997). Traits are observable properties linked to bio-physical or physiological mechanisms that enable plants to cope with environmental stress and/or competition; a PFT (plant functional type) is expected to contain taxa with similar responses to physical or biotic environmental factors, and similar expressions of multiple traits (Harrison et al., 2010). The main problem in adopting a PFT-based approach of bridging the gap between plant physiology and ecosystem processes is how to define the plant functional groupings themselves i.e. how do we select a trait which makes a PFT ‘functional’? Pillar (1999) suggests that the traits must be observable expressions of forms or behaviours defining plant types that are responsive, in terms of occurrence or performance. Plant functional group classifications have been developed for palaeoecological applications based on traits that determine the distribution of plant taxa along climatic gradients with traits relating to influences on soil formation, landscape formation (erosion and sedimentation), catchment hydrology and biogeochemical cycling. Plant functional types (PFTs) are most frequently defined as groups of plants exhibiting either similar responses to an environment or similar effects on major ecosystem processes (Noble and Gitay 1996; Skarpe 1996; Diaz and Cabido 1997; Duckworth et al. 2000; Kleyer 2002; Lavorel and Garnier 2002). According to Walker (1992) the appropriate basis for defining species functional types (often referred to as guilds) is the way the biota regulates ecosystem processes; defining them in this way focuses attention on the processes that maintain ecosystem and community function, and on how changes in the relative or absolute abundances and patterns of distribution of the functional groups concerned will influence these processes. Aspects of Raunkiaer’s (1934) classification (Figure 1.4) underpin many attempts at top-down functional classification of plants (Harrison et al., 2010). The relative dominance of Raunkiaer life forms in plant communities is consistent with generally accepted broad classifications of world vegetation into climatically determined biomes (Smith, 1913, Mueller-Dombois and Ellenberg, 1974). However, precise definitions of PFTs vary, with some researchers favouring a definition based primarily on resource use and Root’s 1967 guild concept (defined as a group of species that exploit the same class of environmental resources in a similar way (Boutin and Keddy, 1993, Wilson, 1999)). Classifying species by shared biological characteristics that relate to function, rather than phylogeny, is increasingly recognised as a tool in identifying common functional response groups among taxa (Grime et al, 1988; Lavorel et al, 1997; Diaz and Cabido, 2001). Different plant functional types are expected to play different roles in terms of matter and energy processes in ecosystems (Pillar, 1999). Under a changing climate, one would expect a shift in the relative abundance of PFTs because of local extinction and migration across the landscape (Huntley, 1991).
Figure 2.22 Raunkiaer (1934) classified plants according to the place where the growth point is located during the less favourable growing seasons.

Research on biodiversity and biological conservation has previously focused on individual species (autecology) (e.g., Perring and Farrell, 1983; Dinsdale et al., 1997) or plant communities described in terms of taxonomically defined species assemblages (synecology) (e.g., Ellenberg, 1988; Rodwell, 1991,1995). However, there has been considerable debate over what constitutes a species (World Conservation Monitoring Centre, 1992; Bisby, 1995; Duarte et al., 1995). Recent studies have highlighted the need to move from life form-based classification to emergent group classification based on functional traits (Epstein et al., 2001; Jeltsch et al., 2008; Harrison et al., 2010; Kattge et al., 2011). This was not possible in the past, due to the problems involved in combining data from regions with taxonomically different floras, but has been now addressed using PFTs (Prentice and Webb, 1998). As there is no agreed global PFT scheme, there are no definitive distribution maps of PFTs (nor global maps of trait values) to compare with climatic maps (Harrison et al., 2010). A functional classification system that provides sufficient detail to capture component processes, but with sufficient abstraction to develop regionally and globally useful models, would meet this need (Pacala & Kinzig, 2002). In the 1990s, the Global Palaeo-vegetation Mapping Project (BIOME 6000) attempted to create fully documented fossil pollen datasets for 6000 ± 500 $^{14}$C yr BP (mid-Holocene) and 18,000 ± 2,000 $^{14}$C yr BP (last glacial maximum) and to reconstruct global maps of vegetation for these time periods (Prentice & Webb, 1998; Prentice et al., 2000), using a standard technique called biomization (Prentice et al., 1996). Palaeoecological records of past vegetation changes provide a valuable resource for the prediction of future changes, with large-scale projects such as the BIOME 6000 project of the International Geosphere-Biosphere Programme (IGBP) (e.g., Prentice and Webb, 1998; Yu et al., 1998) aiming to reconstruct palaeo-vegetation records for 6000 BP in a global synthesis (Duckworth et al., 2000). The PFT concept is a central concept for assigning pollen taxa to biomes (Ni et al., 2010). Early pollen-based biome reconstructions (Yu et al, 1998, 2000) used a very simple implicit classification of PFTs related to the set of PFTs used in the BIOME1 model (Prentice et al., 1992a). When first introduced, this procedure was applicable only to Europe. It has since been used to successfully delineate the major vegetation types of China for the mid-Holocene by Yu et al. (1998, 2000). The classification of PFTs can be hampered by the lack of consistency in the uses of bioclimatic terms among different regions. For example, some 'temperate
forests’ in China have similar physiognomy and climate to forests classified as ‘boreal’ in western Europe or ‘taiga’ in the former Soviet Union (Prentice & Webb, 1998). However, Yu et al. (1998) reclassified tree PFTs and vegetation types in China in order to achieve physiognomic and bioclimatic consistency with the published scheme for Europe. More recently, Ni et al., (2010) successfully used a more explicit, global classification scheme (Harrison et al., 2010) to test the applicability of such a scheme in order to reconstruct changing biomes and palaeo-vegetation records in China during the late Quaternary. However, it is important to note that there is still no generally accepted functional classification of plants.

2.7.5 Assessing changes in plant functional types in the MG1 record

The PFT concept is utilised in this study to examine the changes in plant functional diversity through time in order assess whether plant functional diversity mediates the effects of climate on lake catchment dynamics (including long-term erosion rates). The use of PFTs helps to solve the problem of classifying palaeoecological records (a) by reducing the number of entities considered and (b) by providing an ecological (as opposed to phylogenetic) basis of treating plant data in a compatible way (Prentice et al., 1998).

In this study, plant taxa are classified according to Ni et al.’s (2010) classification scheme and the definition adopted is that a PFT is a group of plants that, irrespective of phylogeny, are similar in a given set of traits and are similar in their association to certain variables (Pillar & Sosinksi, 2003). This global PFT classification is defined explicitly in terms of four traits: life form, leaf form, phenology and bioclimatic tolerances (e.g. cold-tolerance). A major advantage to using this classification scheme is that it includes a comprehensive list of Chinese pollen types which can be used to define each biome. The scheme is based on the current understanding of their significance in terms of adaptation to the physical environment (Harrison et al., 2009; Harrison et al, 2010). The successful reconstructions of Chinese biomes by Ni et al. (2010) based on pollen data and their biomization technique suggests that this new global scheme of PFTs provides a useful basis for vegetation classification and mapping. Thus, this classification scheme was adopted for this study on the basis that it is globally consistent and the classification is explicit in terms of the plant characteristics.

A threshold pollen percentage (>0.5) was defined in order to reduce the noise due to occasional pollen grains derived from contamination or long-distance transport. Pollen taxa with an abundance of >0.5% were allocated to one or more of the PFT classifications on the basis of information about the life form, leaf form, phenology and bioclimatic tolerances of the taxa based on a variety of plant vegetation classification studies carried out in China (Hou, 1983; Walker, 1986; Xiwen & Walker, 1986; Prentice & Webb, 1998; Yu et al., 1998; Jiang et al., 1999; Ni, 2003; Wang & Ni, 2005). The plant functional types are named by four-part codes reflecting the four classification levels in the order bioclimatic tolerance, phenology, leaf form, and life form as per Ni et al. (2010). For example “cool temperate evergreen needle-leaved trees” are c-te.e.n.t (see Table 3.8). A total of 9 PFTs were found to have diagnostic value and were therefore used in the biomization procedure (Figure 3.27 & 3.28).
2.8 Reference climate record

A reference climate record was required in this study, against which the pollen and lake sediment data could be compared through time in order to examine the possible influence that climatic shifts had on plant functional diversity in the region/catchment, and to determine if there is any evidence that changes in the vegetation record (i.e. plant functional diversity) mediate changes in catchment dynamics in the lake over time. In remote areas, such as the Lake Mugecuo catchment, where continuous instrumental meteorological records do not start until the mid-20th century, long-term quantitative climate data from natural archives are essential. Unfortunately, there is currently no published local climate data from natural archives in this area.

A solution to this was to utilise a published speleothem record from Dongge Cave (Dykoski et al., 2005), which provides a continuous record of the EASM over the last 16 ka from δ¹⁸O measurements of stalagmite calcite. This data represents a regional record of changes in the overall pattern of the EASM for Southwest China during the Holocene in the form of δ¹⁸O values.

Calcareous speleothems (stalagmites, stalactites and flowstones) are widespread in karstic environments and grow from dripwaters that degas excess carbon dioxide upon entering caves (Zhang et al, 2011). In carefully chosen sites, speleothems (or stalagmites) can record key aspects of climate variability such as mean annual temperature, rainfall variability, atmospheric circulation changes and vegetation response in a variety of measureable parameters (McDermott, 2004). These measureable parameters include stable isotope ratios, inter-annual thickness variations of growth laminae, growth-rate changes, variations in trace element ratios, organic acid contents and the nature of trapped pollen grains. Speleothems can provide precisely-date records of δ¹⁸O - a geochemical proxy that, in regions dominated by atmospheric convection, is strongly influenced by the amount of rainfall (Dansgaard, 1964). The variability in oxygen isotope ratios collected from speleothems in Chinese caves is often interpreted as a proxy for variability of precipitation, summer precipitation, seasonality of precipitation, and/or the proportion of ¹⁸O to ¹⁶O of annual total rainfall that is related to a strengthening or weakening of the East Asian monsoon (Dayem et al., 2010). Currently, there is a widespread belief that the oxygen isotopes measure some aspect of the strength or intensity of the monsoon, but apparent differences in usage of words such as strength and intensity has complicated interpretations of such isotopic data in terms of variations in climate (Dayem et al., 2010). It has been argued that isotopic variability does not imply differences in precipitation amount; rather it indicates changes in the ratio of summer to winter precipitation, which they refer to as 'monsoon intensity' (e.g., Cai et al., 2006; Cheng et al., 2006, 2009; Dykoski et al., 2005; Kelly et al., 2006; Wang et al., 2008; Yuan et al., 2004).

Therefore, the D₄ δ¹⁸O record from Dongge Cave presented by Dykoski et al. (2005) was used in this study as a guide to the pattern of changes in EAM intensity for the Holocene.

The D₄ δ¹⁸O record is a 16ka stalagmite record which was recovered from Dongge Cave, Guizhou Province in southern inland China (25°17′N, 108°5′E, elevation=680m) and is unaffected by Rayleigh fractionation. If calcite precipitation is rapid during stalagmite growth there can be insufficient time for oxygen isotopes to equilibrate between water and the bicarbonate ion (Hendy, 1971). This leads to Rayleigh fractionation processes in the bicarbonate reservoir and generates systematic increases in δ¹⁸O which have no direct relationship to δ¹⁸O of drip water or to cave temperature (Hu et al., 2008). The Dongge Cave speleothem record shows no sign of kinetic or evaporative effects and can be reliably interpreted as a record of local rainfall composition and temperature (Wang et al., 2005; Dykoski et al., 2005). Similar to Lake Mugecuo,
most of the rainfall (~80%) in the Dongge Cave region falls during the summer monsoon months (May-Oct) (Shi, 2002). The sampling interval of D4 yielded an average time resolution of ~19 years with some portions sampled as high as every 1 to 2 years (Dykoski et al., 2005). Most analytical time-series techniques require input data to be evenly spaced in time. For the purpose of this study, the D4 record was interpolated to an evenly spaced data set and was then transformed using a very low degree of smoothing using the filter function in Matlab v. R2011 which gave a moving average of approximately 66 years (based on the resolution of the pollen record every 2cm= ~66 years.) The purpose of interpolating the data was to get it to the same resolution as the pollen and lake proxy records in order to make them more easily comparable. Some data contain a high-order trend that can be removed by fitting a higher-order polynomial to the data. The polyval function in Matlab v. R2011 was used to remove this type of high-order trend from the climate data (see Appendix A.1.5 for Matlab script).
3) Chapter 3: Results

This chapter presents the results of the fieldwork carried out at Lake Mugecuo in July 2010, and the subsequent analysis of the data obtained there. Lithological and organic analysis, palaeoecological data and cosmogenic $^{10}$Be measurements are presented individually. Results are then further analysed and discussed in a multi-proxy format in discussion Chapters 4, 5 & 6.

All data are plotted against depth in sections 3.1-3.2, and then plotted relative to age thereafter.

3.1 Core stratigraphy

3.1.1 Deep water core MG1

Eleven (Figure 3.1) stratigraphic zones were identified based on the major shifts in the stratigraphy of the MG1 (deep water) core which was extracted from the deepest part of the lake (Figure 2.11). These major shifts are outlined below.

Zone 1-S-1 (383-367cm) is primarily composed of silty clay-rich gyttja. Zone 1-S-2 (367-359cm) consists of a band of light grey clay-rich gyttja which is replaced by clay rich gyttja in Zone S3 (359-342cm). A mixture of silt/clay-rich sediments manifesting as subtle intermittent grey/green banding was observed in Zone 1-S-4 (342-326cm). Zone 1-S-5 (326-299cm) consisted of a silt-clay mixture being followed by band of clay rich gyttja. A return to a mixture of silty clay-rich gyttja is seen in Zone 1-S-6 (299-261cm).

In contrast to the adjacent silt/clay rich sediments, Zone 1-S-7 (261-182cm) is characterised by a series of subtle banding of light grey silt/clay gyttja. Zone 1-S-8 (182-172cm) consists of an intense grey/green band of organic clay-rich gyttja. A transition to light but subtle banding of silt/clay material occurs in Zone 1-S-9 (172-139cm). Zone 1-S-10 (139-130cm) consists of a band of light coloured grey-green banding of clay-rich gyttja which is followed in Zone 1-S-11 (130-0cm) by a substantial period of subtle banding of lightly coloured organic-rich silt/clay sediment.
3.1.2 Shallow water cores MG3 and MG4

Seven major stratigraphic zones (3-S-1-7) were recognised in the MG3 shallow water core (Figure 3.2). Key features include intense periods of very coarse sand banding in the lower section of the core which mark the boundaries of major stratigraphic transition zones.

The basal zone of the MG3 core (3-S-1; 464-445cm) consists of a very coarse sand (16-25mm) sediment matrix. In contrast to adjacent zones, 3-S-2 (445-432cm) shows a transition to a silt/clay rich gyttja.

Zone 3-S-3 (432-392cm) is characterised by coarse sand material. This is replaced by silt/clay rich gyttja marking the start of zone 3-S-4 (392-383cm).

The core stratigraphy shifts once again to a period of extremely coarse 16-25mm) sand (3-S-5) followed by a transition to a dark silt/clay rich gyttja (3-S-6) from 346-324cm.

The upper section of the Mg3 core is characterised by a very organic rich gyttja (324-0cm) containing extremely organic-rich gyttja.
Eight major stratigraphic zones were identified in the MG4 shallow water core (Figure 3.3). Intense banding of coarse sand in the mid- and lower sections characterise this core. These intense periods of coarse sediment banding mark the major stratigraphic zones of the MG4 composite core sequence.

The basal zone (4-S-1; 312-300cm) is characterised by a light brown coloured silt/clay-rich gyttja. This is followed by an intense period (300-269cm) of extremely coarse (>125mm) sand banding. 4-S-3 displays alternating fine sand/silt/clay banding from 269-259cm.

Zone 4-S-4 (259-240) consists of a period of a very coarse (16-25mm) sand sediment matrix. In contrast, in zone 4-S-5 a transition to a very organic-rich gyttja is observed (240-194cm). This is followed by a very sudden shift to a band of very coarse sand from 194-178cm. Zone 4-S-7 exhibits a transition from organic gyttja to another period of coarse sand sediment from 178-170cm.

Sediment composition in the upper section of the core (4-S-8) is primarily composed of organic-rich gyttja with an abundance of plant organics particularly near the top of the core.
3.2 Core chronologies

The results of the chronologies for the MG1, Mg3 and MG4 cores which were produced by linear interpolation between the dated depths are presented here.

3.2.1 Deep water core MG1

12 samples of bulk organic material were processed for AMS $^{14}$C radiocarbon dating to construct an age-depth model for the deep water core MG1 (Figure 3.4). At Lake Mugecuo, one rangefinder date (D12, see Table 3.1) was used to clarify the age of the deepest sediment in the MG1 (deep water) core. Bennet (1994) suggests that radiocarbon ages should be placed evenly throughout a sequence in an effort to minimise errors associated with the dating of specific horizons. Sediment slices with major changes evident in the pollen record were preferentially selected for bulk sediment AMS $^{14}$C radiocarbon dating (Figure 3.4) with several dates used to pin-point the timing of a dramatic crash in Betula in the pollen record in the deepest section of the MG1 core. Dates were then evenly spaced throughout the core.
The dates indicate that the MG1 core extends back to an age of at least 10,650 ± 40 yrs BP i.e. approximately 12.6kcal. yr BP at 381cm (Table 3.1). The MG1 core generally has a relatively low carbon content throughout, indicating that the influence of hard water (an external effect) was minimal (e.g. Bradley, 1999). The age of the sediment at the top of the MG1 core is presumed to be modern day i.e. the year the core was extracted 2010 AD, while the basal date of MG1 was measured at 12.6 kcal.yr BP, giving a mean core resolution of ~33yr cm⁻¹. The analysis undertaken to construct the $P_{Sequence}$ model in OxCal revealed that radiocarbon dates D8, D9 and D11 were statistical outliers (Figure 3.5) and were therefore rejected from the final age-depth model (Figure 3.6). These dates were rejected on the principle that by removing them, the remaining dates in the age-depth model follow a classic power law profile in conjunction with the fact that the δ¹³C values of these dates differ significantly from the δ¹³C values of the dates either side of them. These differences in δ¹³C suggest that these dates are not robust enough to be utilised in the core chronology and thus were disregarded. The age-depth model displays characteristics of self-organization which follows a power law relationship for the composite MG1 core i.e. the relationship between each of the dates suggests that the sediment has been deposited in situ over time in this core sequence. In total, 9 dates were accepted for the final chronology (Table 3.1). The calibrated dates are rounded up to the nearest one hundred years quoted in kcal.yr BP herein. Sedimentation rates between calibrated dates are presented in Table 3.2.
Figure 3.5 OxCal P_Sequence of radiocarbon ages for MG1 deepwater core with outliers included. Horizontal bars represent the 95% confidence intervals of the calibrated ages.
Figure 3.6 OxCal P_Sequence of radiocarbon ages for MG1 deepwater core with outliers excluded. Horizontal bars represent the 95% confidence intervals of the calibrated ages.
Table 3.1 Conventional and calibrated AMS $^{14}$C radiocarbon dating for the MG1 core. Dates denoted with a * and in italics are excluded from the final core chronology. Calibrated dates are quoted to the nearest 100 years.

<table>
<thead>
<tr>
<th>Date Reference</th>
<th>Core Depth (cm)</th>
<th>Median Age (years BP ± 1σ)</th>
<th>Error (±)</th>
<th>Conventional Radiocarbon Age (years BP)</th>
<th>Calibrated Age (cal yr BP)</th>
<th>Midpoint Confidence Limits</th>
<th>$\delta^{13}$C Value</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
<td>50</td>
<td>1780</td>
<td>30</td>
<td>1815</td>
<td>1614</td>
<td>1.7</td>
<td>-26.4</td>
<td>Bulk</td>
</tr>
<tr>
<td>D2</td>
<td>100</td>
<td>2370</td>
<td>30</td>
<td>2651</td>
<td>2336</td>
<td>2.5</td>
<td>-23.2</td>
<td>Bulk</td>
</tr>
<tr>
<td>D3</td>
<td>150</td>
<td>4180</td>
<td>30</td>
<td>4836</td>
<td>4616</td>
<td>4.7</td>
<td>-25.8</td>
<td>Bulk</td>
</tr>
<tr>
<td>D4</td>
<td>200</td>
<td>6180</td>
<td>30</td>
<td>7169</td>
<td>6987</td>
<td>7.1</td>
<td>-27.2</td>
<td>Bulk</td>
</tr>
<tr>
<td>D5</td>
<td>248</td>
<td>8000</td>
<td>40</td>
<td>9009</td>
<td>8718</td>
<td>8.9</td>
<td>-25.9</td>
<td>Bulk</td>
</tr>
<tr>
<td>D6</td>
<td>290</td>
<td>8840</td>
<td>120</td>
<td>10200</td>
<td>9560</td>
<td>9.9</td>
<td>-23.7</td>
<td>Bulk</td>
</tr>
<tr>
<td>D7</td>
<td>322</td>
<td>10271</td>
<td>39</td>
<td>12140</td>
<td>11835</td>
<td>12.0</td>
<td>-22.5</td>
<td>Bulk</td>
</tr>
<tr>
<td>D8</td>
<td>*334</td>
<td>8128</td>
<td>59</td>
<td>9284</td>
<td>8796</td>
<td>9.0</td>
<td>-26.2</td>
<td>Bulk</td>
</tr>
<tr>
<td>D9</td>
<td>*338</td>
<td>9120</td>
<td>40</td>
<td>10404</td>
<td>10207</td>
<td>10.3</td>
<td>-27.3</td>
<td>Bulk</td>
</tr>
<tr>
<td>D10</td>
<td>341</td>
<td>10319</td>
<td>40</td>
<td>12382</td>
<td>11989</td>
<td>12.2</td>
<td>-22.2</td>
<td>Bulk</td>
</tr>
<tr>
<td>D11</td>
<td>*343</td>
<td>9392</td>
<td>42</td>
<td>10721</td>
<td>10512</td>
<td>10.6</td>
<td>-28.6</td>
<td>Bulk</td>
</tr>
<tr>
<td>D12</td>
<td>381</td>
<td>10650</td>
<td>40</td>
<td>12683</td>
<td>12543</td>
<td>12.6</td>
<td>-22.8</td>
<td>Bulk</td>
</tr>
</tbody>
</table>

Table 3.2 Sedimentation rates between dated points for MG1 core.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Interval (cm)</th>
<th>Median Age (cal yr BP)</th>
<th>Interval (yr)</th>
<th>Average Sedimentation Rate Between Points (yr cm$^{-1}$)</th>
<th>Sedimentation Rate (cm yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>50</td>
<td>0-50</td>
<td>1715</td>
<td>1715</td>
<td>34.3</td>
<td>0.03</td>
</tr>
<tr>
<td>100</td>
<td>50-100</td>
<td>2494</td>
<td>779</td>
<td>15.58</td>
<td>0.06</td>
</tr>
<tr>
<td>150</td>
<td>100-150</td>
<td>4726</td>
<td>2232</td>
<td>44.64</td>
<td>0.02</td>
</tr>
<tr>
<td>200</td>
<td>150-200</td>
<td>7078</td>
<td>2352</td>
<td>47.04</td>
<td>0.21</td>
</tr>
<tr>
<td>248</td>
<td>200-248</td>
<td>8864</td>
<td>1786</td>
<td>37.2</td>
<td>0.03</td>
</tr>
<tr>
<td>290</td>
<td>248-290</td>
<td>9880</td>
<td>1016</td>
<td>24.19</td>
<td>0.04</td>
</tr>
<tr>
<td>322</td>
<td>290-322</td>
<td>11988</td>
<td>2108</td>
<td>68</td>
<td>0.02</td>
</tr>
<tr>
<td>341</td>
<td>322-341</td>
<td>12186</td>
<td>198</td>
<td>9.9</td>
<td>0.10</td>
</tr>
<tr>
<td>381</td>
<td>341-381</td>
<td>12613</td>
<td>427</td>
<td>10.68</td>
<td>0.09</td>
</tr>
</tbody>
</table>
### 3.2.2 Shallow water core MG3

3 samples of bulk organic material were processed for AMS $^{14}$C radiocarbon dating to construct an age-depth model for the shallow water core MG3 (Figure 3.7). The selection of material for AMS $^{14}$C dating was based on the need to constrain the ages of the bands of sand sediment that were used for $^{10}$Be erosion-rate measurements in the core (Figure 3.7). The selection of samples for dating was based upon the composite core stratigraphy, particle size data, and the availability of dateable material. Several attempts were made to use a pollen dating technique but the sediment adjacent to the units of sand was simply too coarse and organic-poor to archive the required quantity of pollen necessary for $^{14}$C analysis.

![Figure 3.7 Selection of material for AMS $^{14}$C radiocarbon dating in the MG3 core. Sections of the core highlighted in red represent the material used for $^{10}$Be erosion rate measurements. Mid-point calibrated ages are highlighted in red and italics to the left. Particle size data is included to illustrate how the $^{14}$C dates are used to constrain the ages of the bands of sand sediment that were used for $^{10}$Be erosion-rate measurements. The date denoted with a * is excluded from the final core chronology.](image)

The dates indicate that the MG3 core extends back to an age of at least 3127 ± 28 years BP i.e. approximately 3.4kcal. yr BP at 381cm (Table 3.3). The 3 $^{14}$C dates were calibrated and tested for internal consistency using Oxcal v. 4.17 and Intcal09 (Bronk Ramsey, 1995; Bronk Ramsey, 2008; Reimer et al., 2009). In total, 2 dates were accepted for the final chronology (Table 3.3, Figure 3.8). Due to the age-reversal of the SW-1 (433cm) date, this date was not accepted. The calibrated dates are rounded up to the nearest one hundred years quoted in kcal.yr BP herein. Calculated sedimentation rates are indicative only (Table 3.4) as they assume a constant rate of deposition between calibrated dates in order to infer key shifts in sedimentation rates. Figure 3.8 shows the approximate number of years represented by each cm of sediment in the MG3 core.
Figure 3.8 *OxCal P_Sequence* of radiocarbon ages for the MG3 shallow water core. Horizontal bars represent the 95% confidence intervals of the calibrated ages.

Table 3.3 Conventional and calibrated AMS 14C radiocarbon dating for the MG3 shallow water core.Calibrated dates are quoted to the nearest 100 years. Dates denoted with a * and in italics are excluded from the final core chronology.

<table>
<thead>
<tr>
<th>Date Reference</th>
<th>Median Core Depth (cm)</th>
<th>Conventional Radiocarbon Age (years BP ± 1σ)</th>
<th>Error (±)</th>
<th>Calibrated Age (cal.yr BP)</th>
<th>95% Confidence Limits</th>
<th>Mid-point (kcal.yr BP)</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>SW6</td>
<td>343</td>
<td>2688</td>
<td>28</td>
<td>2753</td>
<td>2847</td>
<td>2.8</td>
<td>Bulk</td>
</tr>
<tr>
<td>SW7</td>
<td>381</td>
<td>3127</td>
<td>28</td>
<td>3265</td>
<td>3440</td>
<td>3.4</td>
<td>Bulk</td>
</tr>
<tr>
<td>SW1</td>
<td>*433</td>
<td>2404</td>
<td>38</td>
<td>2698</td>
<td>2343</td>
<td>2.5</td>
<td>Bulk</td>
</tr>
</tbody>
</table>
Table 3.4. Sedimentation rates between dated points for MG3 core.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Interval (cm)</th>
<th>Median Age (cal yr BP)</th>
<th>Interval (yr)</th>
<th>Average Sedimentation Rate Between Points (yrs/cm)</th>
<th>Sedimentation Rate (cm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>343</td>
<td>0-343</td>
<td>2800</td>
<td>2800</td>
<td>8.16</td>
<td>0.12</td>
</tr>
<tr>
<td>381</td>
<td>343-381</td>
<td>3353</td>
<td>553</td>
<td>14.55</td>
<td>0.07</td>
</tr>
</tbody>
</table>

3.2.3 Shallow water core MG4

4 samples of bulk organic material were processed for AMS $^{14}$C radiocarbon dating to construct an age-depth model for the shallow water core MG4 (Figure 3.9). The selection of material for AMS $^{14}$C dating was based on the need to constrain the ages of the bands of sand sediment that were used for $^{10}$Be erosion-rate measurements in the core (Figure 3.9). The selection of samples for dating was based upon the composite core stratigraphy, particle size data, and the availability of dateable material. Several attempts were made to use a pollen dating technique but the sediment adjacent to the units of sand was simply too coarse and organic-poor to archive the required quantity of pollen necessary for $^{14}$C analysis.

The dates indicate that the MG4 core extends back to an age of at least 1375 ± 27 years BP i.e. approximately 1.3kcal. yr BP at 303cm (Table 3.5). The 4 $^{14}$C dates were calibrated and tested for internal consistency using Oxcal v. 4.17 and Intcal09 (Bronk Ramsey, 1995; Bronk Ramsey, 2008; Reimer et al., 2009). In total, 4 dates were accepted for the final chronology (Table 3.5; Figure 3.10). The calibrated dates
are rounded up to the nearest one hundred years quoted in kcal.yr BP herein. Calculated sedimentation rates are indicative only (Table 3.6) as they assume a constant rate of deposition between calibrated dates in order to infer key shifts in sedimentation rates. Figure 3.9 shows the approximate number of years represented by each cm of sediment in the MG4 core.

Figure 3.10 OxCal P_Sequence of radiocarbon ages for the MG4 shallow water core. Horizontal bars represent the 95% confidence intervals of the calibrated ages.

Table 3.5 Conventional and calibrated AMS $^{14}$C radiocarbon dating for the MG4 shallow water core. Calibrated dates are quoted to the nearest 100 years.

<table>
<thead>
<tr>
<th>Date Reference</th>
<th>Median Core Depth (cm)</th>
<th>Conventional Radiocarbon Age (years BP ± 1σ)</th>
<th>Error (±)</th>
<th>Calibrated Age (cal.yr BP) 95% Confidence Limits</th>
<th>Mid-point (kcal.yr BP)</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>SW2</td>
<td>185</td>
<td>755</td>
<td>27</td>
<td>667</td>
<td>728</td>
<td>0.7</td>
</tr>
<tr>
<td>SW3</td>
<td>193</td>
<td>822</td>
<td>27</td>
<td>686</td>
<td>784</td>
<td>0.7</td>
</tr>
<tr>
<td>SW4</td>
<td>255</td>
<td>1176</td>
<td>27</td>
<td>1000</td>
<td>1176</td>
<td>1.1</td>
</tr>
<tr>
<td>SW5</td>
<td>303</td>
<td>1375</td>
<td>27</td>
<td>1270</td>
<td>1338</td>
<td>1.3</td>
</tr>
</tbody>
</table>
Table 3.6 Sedimentation rates between dated points for MG4 core.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Interval (cm)</th>
<th>Median Age (cal.yr BP)</th>
<th>Interval (yr)</th>
<th>Average Sedimentation Rate Between Points (yrs/cm)</th>
<th>Sedimentation Rate (cm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>185</td>
<td>0-185</td>
<td>698</td>
<td>698</td>
<td>3.77</td>
<td>0.27</td>
</tr>
<tr>
<td>193</td>
<td>185-193</td>
<td>735</td>
<td>37</td>
<td>4.63</td>
<td>0.22</td>
</tr>
<tr>
<td>255</td>
<td>193-255</td>
<td>1088</td>
<td>353</td>
<td>5.70</td>
<td>0.18</td>
</tr>
<tr>
<td>303</td>
<td>255-303</td>
<td>1304</td>
<td>216</td>
<td>4.50</td>
<td>0.22</td>
</tr>
</tbody>
</table>

3.2.4 Unified core chronologies

Figures 3.11 & 3.12 present the unification of the core chronologies from the MG1, MG3 and MG4 cores. The age-depth models for each of the cores were combined to relate the erosion-rates measured in the shallow cores to dated sediment in the deep water core. As outlined previously it was not possible to simply directly compare $^{10}$Be erosion rates calculated from two shallow water cores with pollen concentrations in the deep water core; to draw this comparison it was necessary to correlate the shallow water core chronologies with the deep water core chronology.

![Figure 3.11](image)

Figure 3.11. The panel on the left represents the established age-depth model for the MG1 deep water core. The panels on the right highlight the concurrent time periods covered by the shallow water core chronologies.
Figure 3.12. The panel on the left shows the age-depth model for the deep water core (MG1) established from 9 AMS $^{14}$C bulk sediment dates. The panel on the right is an excerpt that highlights the uppermost 130cm which unifies the core chronologies of the shallow water cored (MG3 & MG4) with the chronology of the MG1 core.

The ages of the palaeo-deposits of sand used for $^{10}$Be analysis are presented herein as the mid-point calibrated age of the sand units based on this new unified core chronology. The approximate ages at the boundaries of each of the bands were used to constrain these ages (Table 3.7). The age of the MG3c sample was constrained by interpolating the age-depth profile constructed using the two accepted dates from the MG3 core chronology (see section 3.2.2).

Table 3.7 Constrained ages of the palaeo-deposits of sand based on a unified core chronology with MG1.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Sample Age Range kcal.yr BP</th>
<th>Mid-point Sample Age kcal.yr BP</th>
<th>Mid-point Sample Depth (cm) (Based on Unified Core Chronology)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MG3a</td>
<td>2.8-3.4</td>
<td>3.0</td>
<td>112</td>
</tr>
<tr>
<td>MG3b</td>
<td>3.4-3.5</td>
<td>3.4</td>
<td>120</td>
</tr>
<tr>
<td>MG3c</td>
<td>3.5-4.1</td>
<td>3.8</td>
<td>130</td>
</tr>
<tr>
<td>MG4a</td>
<td>0.7-0.7</td>
<td>0.7</td>
<td>21</td>
</tr>
<tr>
<td>MG4b</td>
<td>1.1-1.3</td>
<td>1.2</td>
<td>35</td>
</tr>
</tbody>
</table>
3.3 Lithological and Organic Analysis

The proxy curves of the particle size, loss-on-ignition, total organic carbon, total organic nitrogen, carbon/nitrogen & $\delta^{13}$C analysis on the MG1 core are presented here.

3.3.1 Particle size

Figure 3.13 shows the relative percentages of sand, silt and clay detected in the MG1 core. Particle size data indicates that the core is primarily composed of around 55-85% silt (Fig. 3.13), with some increased levels of organic gyttja at 320-305cm, 180-175cm, 148-128cm. Clay values range from ~15-35%. Sand content is generally low <3.5%. The particle size record is divided into 4 zones (PS1-4) (Figure 3.13), delineated on the basis of prominent shifts in the record.

Zone PS1 (12.6-10.4 kcal.yr BP; 383-298cm) is primarily composed of silt (~75%) and clay (~25%). Sand content is low (< 1%). A marked decline in silt is seen from 320-303cm while clay increases to highs of ~30% for the same period.

Zone PS2 (10.4-6.3 kcal.yr BP; 298-184cm) commences with a rise in silt levels to ~80%. This is followed by increasing levels in silt for the remainder of the zone until they reach their highest levels for the entire MG1 core of almost 95%. The lowest contributions of clay to the whole core are recorded at 8% at 195cm. Sand levels reach their highest of the core of 3% at 270cm.

Zone PS3 (6.3-3.1 kcal.yr BP; 184-113cm) sees a dramatic drop in silt levels to 58-60% from 184-171cm, the lowest values recorded in the MG1 core. Clay contributions reach their highest of the entire core (>35%) from 184-171cm. Sand content remains low.

Zone PS4 (3.1-0 kcal.yr BP; 113-0cm) contains relatively stable levels of silt (~80%) and clay (15-20%) content throughout. For the first time in the record a sustained period of increased sand content occurs from 68-35cm. Levels of sand reach highs of ~2.5% during this period.
3.3.2 % Loss-on-ignition

%LOI data indicates that levels of organic matter in the lower section of the core (12.6-9.9 kcal.yr BP) range from ~15-22% but are noticeably lower (~5-10%) in the upper section of the core (9.9-0.0 kcal.yr BP). Three LOI zones are delineated based on the major shifts in the proxy curve (Figure 3.14).

**Zone LOI-1** (12.6-11.0 kcal.yr BP; 383-308cm) exhibits the highest organic content of the core (>20%) in the form of three separate peaks at 12.5, 12.4 and 12.1 kcal.yr BP.

**Zone LOI-2** (11.0-7.8 kcal.yr BP; 308-220cm) reveals significant fluctuations in the organic profile. %LOI values range from lows of <5% at 11.0 kcal.yr BP and 9.2 kcal.yr BP to highs of >15% at 10.2, 9.1-8.9 and 8.4 kcal.yr BP.

**Zone LOI-3** (7.8-0.0 kcal.yr BP; 220-0cm) is characterised by fluctuating organic contents ranging from 5-15% which indicate a downward shift in the levels of organic matter compared to earlier sections of the core.
3.3.3 Total organic carbon, total organic nitrogen and C/N

The results of total-organic carbon analysis, total nitrogen analysis and C/N analysis are presented in Figure 3.15. Four prominent zones were determined based on major shifts in the proxy curves. %TOC values range from 0.5% to 9.5%. %TN values are generally low throughout the MG1 core, <0.8%. C/N values as a result are relatively low throughout the core ranging from 7.5-13.5.

**Zone OM-1** (12.6-9.3 kcal.yr BP; 383-268cm) reveals fluctuations in %TOC with lows of 3.5-4.5% and highs of 8.5-9.5%. %TOC levels remain stable from approximately 310-268cm i.e. 11.2-9.3 kcal.yr BP. This zone is characterised by variable total-organic nitrogen values ranging from 0.3-0.8%. %TN values stabilize from approximately 11.2-9.3 kcal.yr BP with values ranging from 0.6-0.7%. C/N values reach lows of ~8 in this section of the core before recovering to an average of around 11.5.

**Zone OM-2** (9.3-6.5 kcal.yr BP; 268-187cm) sees %TOC peak at ~9% followed by a dramatic drop to ~4.5%. Organic carbon levels then shift towards declining trend for the remainder of this zone. This zone reveals
more stable %TN values averaging ~0.5%. %TN values begin to drop at approximately 8.6 kcal.yr BP averaging between 0.3-0.5%. The record indicates relatively stable mean C/N ratio values of 11.5-13.5.

**Zone OM-3** (6.5-2.3 kcal.yr BP; 187-87cm) sees a dramatic drop in %TOC to their lowest values of the whole core at 0.5%. Levels recover to values of ~6.5% but drop again from 145-130cm (from 6.5% to a low of 2.5%). This zone begins with a dramatic drop in nitrogen content with the lowest values of the whole core (less than 0.1%). Values recover to nearly 0.6% but subsequently drop again to 0.2% before recovering to average ~0.4% for the remainder of the zone. This zone is characterised by variable C/N values ranging from 10-14 culminating in a peak at 2.3 kcal.yr BP.

**Zone OM-4** (2.3-0.0 kcal.yr BP; 87-0cm) reveals dramatic fluctuations in %TOC values ranging from 3-8%. A relatively stable period of total organic carbon content occurs from 1.5-0.7 kcal.yr BP but from this section to the top of the core values fluctuate once again. This zone is characterised by variable levels of %TN ranging from 0.3-0.6% and more stable C/N values before a drop occurs at approximately 1.3kcal.yr BP to ~9.5. Values recover following this drop, ranging from 10.5-11.5 for the remainder of the core.

Figure 3.15 C/N ratios with major changes in the proxy curves used to delineate the record into zones.
3.3.4 $\delta^{13}C$

$\delta^{13}C$ values are presented in Figure 3.16. $\delta^{13}C$ values for the MG1 core range from -28 to -18‰. $\delta^{13}C$ values peak at 11.4 kcal.yr BP (with a value of -18‰) but steadily decline where they remain between -24‰ and -26‰ throughout the upper section of the record. The $\delta^{13}C$ record is divided into six zones on the basis of the major shifts evident in the proxy curve.

**Zone DC-1** (12.6-12.2 kcal.yr BP; 383-341cm) is characterised by some dramatic shifts in the $\delta^{13}C$ record. Values range from -24 to -19‰ with major peaks at 12.5 and 12.3 kcal.yr BP.

**Zone DC-2** (12.2-10.0 kcal.yr BP; 341-292cm) in contrast, begins with notably lower $\delta^{13}C$ values ranging from -27 to -24‰ before peaking at approximately 11.4 kcal.yr BP) with a value of -19‰.

The beginning of **Zone DC-3** (10.0-9.0 kcal.yr BP; 292-252cm) displays relatively stable $\delta^{13}C$ of approximately -25‰, before a dramatic shift to lows of -27‰.

**Zone DC-4** (9.0-6.6 kcal.yr BP; 252-189cm) presents fluctuations in $\delta^{13}C$ values between -28 to -26‰.

**Zone DC-5** (6.6-3.9 kcal.yr BP; 189-131cm) sees a general increasing trend in $\delta^{13}C$ values i.e. less negative values.

**Zone DC-6** (3.9-0.0 kcal.yr BP; 131-0cm) is characterised by a slight increase in $\delta^{13}C$ values which average approximately -26‰.

![Figure 3.16 $\delta^{13}C$ record with core stratigraphy and major zones.](image)
3.3.4.1 Combined results of $\delta^{13}$C and C/N analysis

The cross-plotting of the $\delta^{13}$C values and C/N ratios as per Meyers et al. (1999) revealed the following trends (Figure 3.17). If a sample has a low C/N ratio and more negative $\delta^{13}$C values the organic material is most likely derived from low productivity aquatic sources (Meyers et al., 1997). The majority of the samples analysed fall into the lacustrine plants category.

![Figure 3.17. MG 1 C/N ratios plotted against $\delta^{13}$C values quoted in ‰ (per mille). Fields for the different plant types are from Meyers et al., 1999.](image)

3.3.5 X-ray diffraction analysis

The clay mineral assemblages of the samples tested by X-ray diffraction analysis (Table 3.8) were dominated by high levels of illite (70-90 %). Minor and variable amounts of kaolinite, chlorite and smectite were also present in these samples.

<table>
<thead>
<tr>
<th>Sample</th>
<th>%clay mineral</th>
<th>Non-clay minerals</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>illite</td>
<td>kaolinite</td>
</tr>
<tr>
<td>Modern Clay Sample retrieved from lake</td>
<td>94</td>
<td>4</td>
</tr>
<tr>
<td>Mg1 (deep water core) 172-182cm</td>
<td>82</td>
<td>10</td>
</tr>
<tr>
<td>Mg1 (deep water core) 320-299cm</td>
<td>84</td>
<td>5</td>
</tr>
</tbody>
</table>
3.4 Pollen analysis

3.4.1 Basic analysis versus age

China supports a vast range of plant taxa and is one of the most ecologically diverse countries in the world making it very challenging to interpret the pollen record due to the vast number and type of species represented. The pollen sum (including trees, shrub, and fern spores) was comprised of ~400 pollen grains per sample. A total of 95 pollen taxa were distinguished but only those that appear at least 3 times in the record with percentages above 0.1% or were consistent in more than ten samples are pictured in the pollen diagrams due to the large volume of taxa identified. Pollen concentration (grains/g) and pollen percentage diagrams are presented in Figures 3.18 to 3.20 outlining the changes in arboreal and non-arboreal taxon over time. The arboreal pollen mainly consists of Betula, Pinus, Carpinus, Cyclobalanopsis, Taxodiaceae, Tsuga, Lithocarpus and Quercus. Non-arboreal contributions come from Rosaceae, Artemisia, Ericaceae, Euphorbiaceae, Cyperaceae and Poaceae. Figure 3.20 highlights the total abundances of arboreal, non-arboreal and fern pollen present in the core along with total pollen concentration.

Betula and Pinus pollen feature strongly throughout the pollen record. Broadly speaking, the lower section of the MG1 record (12.6-9.4 kcal.yr BP; 383-270cm) is dominated by Betula with contributions from Pinus, Carpinus, Artemisia and Cyperaceae. Conversely, the upper section of the record (9.4-0.0 kcal.yr BP; 270-0cm) is dominated by Pinus with contributions from Betula, Cyclobalanopsis, Taxodiaceae, Rosaceae, Artemisia, Poaceae and Cyperaceae. Aquatics are low throughout the core. Total pollen concentrations do not fluctuate heavily throughout the record with values generally ranging from <30,000 grains/g to highs of almost 60,000 grains g⁻¹.
Figure 3.18 MG1 Pollen percentages of arboreal (AP) and non-arboreal (NAP) taxa including trees, shrubs, herbs, grasses and sedges. Spores/ferns and aquatics are also included. Due to the large number of species identified, taxa with an abundance <0.1% are excluded from this diagram. Pollen zones (derived using CONISS including taxa with abundances of >0.1%) and core stratigraphy are also included for reference.
Figure 3.19 Pollen concentration (grains/g) of arboreal (AP) taxa non-arboreal (NAP) taxa including trees, shrubs, herbs, grasses and sedges. Spores/ferns and aquatics are also included. *Betula* and *Pinus* are quoted as $10^3$. *Cyclobalanopsis*, *Taxodiaceae*, *Carpinus*, *Tsuga*, *Quercus*, *Artemisia*, *Rosaceae*, *Euphorbiaceae*, *Poaceae* and *Cyperaceae* are quoted as $10^2$. Pollen zones (derived using CONISS including taxa with abundances of $>0.1\%$), and core stratigraphy are also included for reference.
3.4.2 Pollen zones

Distinct pollen assemblage zones were defined using constrained incremental sum-of-squares cluster analysis (CONISS; Grimm, 1987), based on the square-root transformed percentages of all pollen types in the pollen diagram (Figure 3.18).

Zone P-1 (12.6-11.9 kcal yr BP; 383-322 cm) spans approximately 700 years. *Betula* is the primary contributor to the pollen record in this zone accounting for 65-80% of the pollen. *Pinus* accounts for 17-32% while *Artemisia* accounts for approximately 7-9%. Cyperaceae levels were initially high, accounting for ~10% but declining to ~4%. Contributions also come from, *Carpinus*, *Cyclobalanopsis*, *Quercus*, Cyperaceae, and *Thalictrum*. Total pollen concentrations (TPC) reach highs of ~60,000 grains/g in this section of the record. Increasing levels of arboreal pollen (AP) are seen reaching highs of almost 50,000 grains/g. Arboreal pollen counts for over 80% of the total pollen percentages during this period. AP concentrations range from 20,000 to almost 50,000 grains/g.

*Betula* has average values of approximately 10,000-20,000 grains/g making it the most abundant taxon. *Pinus* values fluctuate considerably in this zone ranging from ~5% up to highs of 50%. *Pinus* concentration is typically ~5,000 grains/g. *Carpinus* pollen is the next highest arboreal contributor increasing from
approximately 1,000 grains/g to just over 3,000 grains/g. * Artemisia* is the highest non-arboreal contributor in this zone with levels increasing from 2,500 grains/g to approximately 4,500 grains/g. Cyperaceae abundance rates were initially high ~3,000 grains/g but decline to levels of approximately 2,000 grains/g.

This section of the MG1 record includes a pronounced shift in the palaeoecological record ca. 12.3 kcal.yr BP which is one of the most prominent features of the MG1 record. This zone is characterised by high levels of *Betula* (50-60%) which abruptly decline from 50% at 12.3 kcal.yr BP (347cm) to lows of 23% at 12.0 kcal.yr BP (329cm) spanning a period of ca. 300 years. During this period *Betula* pollen concentrations fall from approximately 30,000 grains/g to < 15,000 grains/g. *Pinus* then becomes the primary contributor to the pollen record following the decline in *Betula*. Evidence that this dramatic shift is not just an anomaly in the record can be seen in the increased levels of *Cyclobalanopsis*, *Taxodiaceae*, *Tsuga* and Cyperaceae which occur during this period.

Notable contributions also come from fluctuating concentrations of *Carpinus* pollen (ranging from ~1,000 up to almost 4,000 grains/g) and *Taxodiaceae* with levels increasing up to highs of almost 10% or 3,000 grains/g. In terms of non-arboreal taxa, *Artemisia* levels go into decline in this section of the record from highs of 4,500 grains/g to lows of 1,000 grains/g. Poaceae levels remain low in this zone (<1,000 grains/g), while Cyperaceae is relatively stable until approximately 12.0 kcal.yr BP where a peak in values is seen (~4,000 grains/g).

**Zone P-2** (11.9-8.3 kcal.yr BP; 322-232cm) spans approximately 3,600 years. *Betula* recovers in this zone to reach a high of ~65%. A notable decline in *Betula* follows this high suggesting a drop in local and/or regional birch pollen production. As *Betula* goes into decline during this phase, *Pinus* is on the incline reaching highs of ~40%.

Arboreal pollen accounts for approximately 80-85% of the total pollen percentages during this period. A notable peak in AP pollen concentrations is seen at 9.4 kcal.yr BP (270cm) which corresponds to a peak in *Quercus* pollen concentration, its highest of the entire record, at nearly 3,000 grains/g. Although concentrations are lower than previous periods, *Carpinus* pollen continues to be a noteworthy contributor to the total pollen concentrations of the record with levels ranging between 1,000-2,500 grains/g. *Artemisia*, Rosaceae, Poaceae and Cyperaceae are the primary non-arboreal contributors to this section of the record, maintaining relatively stable pollen concentrations.

**Zone P-3** (8.3-5.8 kcal.yr BP; 232-172cm) spans 2,500 years and is characterised by a further decline in *Betula* pollen and an increase in *Pinus* which becomes the primary contributor to the record during this time. *Pinus* increases to highs of 30-40% while *Betula* reaches lows of 15-30%; its lowest levels for the entire record so far. Both *Betula* and *Pinus* present average pollen concentrations of approximately 10,000 grains with some notable peaks in *Pinus* of up to 20,000 grains in the upper section of this zone. Notable arboreal contributors include *Cyclobalanopsis*, *Taxodiaceae*, *Carpinus* and *Tsuga*. TPC levels recover slightly from the lows of the previous zone ranging between 25,000 and 40,000 grains/g. A peak in arboreal pollen percentage and concentration is attributed to a peak in *Carpinus* pollen at approximately 6.7 kcal.yr BP (192cm). A notable decline in the record from approximately 6.1-5.8 kcal.yr BP of arboreal pollen coincides with an increase in Rosaceae pollen from ~1,000 to 3,000 grains/g which is the primary contributor to NAP.
during this period. Non-arboreal pollen contributions to the record are on the rise in this zone reaching highs of just over 30%.

**Zone P-4** (5.8-2.3 kcal.yr BP; 172-85cm) spans ca. 3,500 years. Decreased levels of *Betula* continue throughout this zone reaching lows of <10%. A marked shift resulting in increased contributions from *Cyclobalanopsis*, Taxodiaceae, *Carpinus* and *Tsuga* occur between approximately 5.0-3.5 kcal.yr BP (155-123cm). Following this shift, from approximately 3.5 to 2.2 kcal.yr BP (123-84cm), *Pinus* pollen increases to highs of ~50-60% with pollen concentrations of approximately 15,000-20,000 grains/g. Declines in *Cyclobalanopsis*, Taxodiaceae, *Carpinus* are noteworthy at this point, while *Tsuga* maintains relatively high pollen concentrations ranging from 1,000-3,000 grains/g.

TPC remain relatively stable during this period. AP% fluctuates in this zone contributing to ~68-80% of the pollen in the record. NAP concentrations also fluctuate in this zone with a marked shift to declining pollen concentrations from approximately 3.3-2.5 kcal.yr BP (119-100cm).

**Zone P-5** (2.3-0 kcal.yr BP; 85-0cm) spanning approximately 2,300 years is characterised by a decline in *Pinus* pollen and a notable increase in NAP %. *Pinus* displays a downward trend in percentage and concentration levels in this zone to lows of <20% and < 10,000 grain/g. Although AP% are generally in decline during this period, marked increases in taxa such as *Cyclobalanopsis*, Taxodiaceae are observed. In terms of NAP%, marked increases in *Artemisia*, Rosaceae, Poaceae and Cyperaceae contribute to the increased presence of non-arboreal pollen taxa in the record. A notable peak in Euphorbiaceae pollen occurs from approximately 1.7-1.2 kcal.yr BP (55-35cm) with pollen concentrations ranging from 1,200 to 2,000 grains/g, its highest concentration of the entire record. Total pollen concentrations generally remain relatively stable.

### 3.4.3 DCA/PCA analysis

The results of the numerical analyses (Detrended Canonical Analysis & Principal Components Analysis) are presented below and illustrated in Figure 3.21. Fifteen pollen taxa with a maximum abundance of >0.5% and at least 3 occurrences in 192 samples were included in the data set. The DCA analysis provides eigenvalues of 0.053 for the first axis and 0.015 for the second axis. The results of the DCA analysis pointed to a linear response of species data to environmental gradients (Birks, 1998). These environmental gradients were interpreted based on the PCA results.

The PCA analysis results are presented in Figure 3.22. Samples clustered together denoted by a circle symbol represent samples which fall between 0-87cm, while samples clustered together denoted by a square symbol represent samples which fall between 88-383cm. The first principal component (Axis 1) captures 32% of the total variance of the dataset. The second principal component (Axis 2) captures 18% of the total variance indicating that both axes together represent 50% of the variance of the total dataset.
Figure 3.21. DCA ordination of the pollen percentage data showing sample distributions in the first two axes.
Figure 3.22 PCA scattergraph showing the clustering of samples from 0-383m from MG1. The samples are split into two clusters. The first cluster (circles) represents samples from 88-383cm. The second cluster (squares) represents samples from 0-87cm.
3.4.4 Pollen diversity/evenness

The results of the Shannon-Wiener analysis, Pielou Evenness and across-sample rarefaction are presented in Figure 3.23.

The H' values for the MG1 core average between 1.5-2, peaking at 2.25 in the upper section of the core and falling below 1.5 from ~127-100cm. Pielou's evenness is a measure of equitability which compares the observed Shannon-Wiener index against the distribution of individuals between the observed species which would maximise diversity i.e. it is a measure of the contribution of species number and equitability to changes in diversity. Pielou's evenness fluctuates between values of 0.5-0.9 in the MG1 core. Across-sample rarefaction estimates the species richness of every sample in a data set for a standard number of individuals i.e. it estimates the species richness of each sample if they contained the same number of individuals. This allows for different sampling efforts when the total number of individuals counted in each sample varies from sample to sample. Figure 3.23. indicates that the species richness does not vary radically between each sample; values range from 6-11.

![Figure 3.23](image)

Figure 3.23 Summary diagram of Pielou’s Evenness, Shannon-Weiner Diversity Index (H') and across-sample rarefaction versus age. Pollen zones (PS1-5) are also included for reference.

3.4.5 Climatic Indicators

The *Artemisia*/Cyperaceae (A/Cy) ratio results are presented in Figure 3.24. The values of A/Cy are generally <10. The record peaks in several places, most notably at 12.6 kcal.yrBP (378cm) where it peaks at 25 and a period of high values from approximately 9.4-8.9 kcal.yr BP (270-250cm) peaking at 18. The upper
section of the core also sees a period of higher values from approximately 4.8-2.2 kcal.yr BP (151-79cm) ranging from values of 5 to 15.

The arboreal/non-arboreal pollen ratio results are presented below. Ratios are generally slightly higher (~5) in the lower section of the record from approximately 12.6-6.6 kcal.yr BP (383-190cm) but drop slightly in the upper section from 6.6-0.0 kcal.yr BP (190-0cm) to around values <5.

3.4.6 Plant functional classification

The results of the plant functional classification are presented in Tables 3.9-3.10 and Figures 3.25-3.26. Nine plant functional types (PFTs) are recognised based on this classification. Each of these nine PFTs is assigned to a biome as per Ni et al. (2010) utilising the nomenclature used in BIOME4 (Kaplan, 2001; Kaplan et al., 2003) and aspects of the classification used in BIOME1 (Prentice et al., 1992a). The main changes in the plant functional classification data are in parallel to the major changes in the main pollen record so are zoned and presented accordingly.
Table 3.9 Classification of pollen taxa on the basis of a global PFT classification as per Ni et al. (2010)

<table>
<thead>
<tr>
<th>Plant Taxa/Family</th>
<th>Life Form</th>
<th>Leaf Form</th>
<th>Phenology</th>
<th>Bioclimatic Tolerance</th>
<th>Plant Functional Classification</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pinus</td>
<td>Tree</td>
<td>Needle-Leaved</td>
<td>Evergreen</td>
<td>Cool-Temperate</td>
<td>Cool-Temperate Evergreen Needle-Leaved Tree</td>
</tr>
<tr>
<td>Tsuga</td>
<td>Tree</td>
<td>Needle-Leaved</td>
<td>Evergreen</td>
<td>Cool-Temperate</td>
<td>Cool-Temperate Evergreen Needle-Leaved Tree</td>
</tr>
<tr>
<td>Ericaceae</td>
<td>Shrub</td>
<td>Small-Leaved</td>
<td>Evergreen</td>
<td>Temperate</td>
<td>Drought-Intolerant Small-Leaved Low Or High Shrub</td>
</tr>
<tr>
<td>Poaceae</td>
<td>Grass</td>
<td></td>
<td></td>
<td></td>
<td>Grass Graminoid</td>
</tr>
<tr>
<td>Cyperaceae</td>
<td>Sedge</td>
<td></td>
<td></td>
<td></td>
<td>Sedge Graminoid</td>
</tr>
<tr>
<td>Carpinus</td>
<td>Tree</td>
<td>Broad-Leaved</td>
<td>Cold-Deciduous</td>
<td>Temperate (Spring-Frost Tolerant)</td>
<td>Temperate (Spring-Frost Avoiding) Cold-Deciduous Malacophyll Broad-Leaved Tree</td>
</tr>
<tr>
<td>Betula</td>
<td>Tree</td>
<td>Broad-Leaved</td>
<td>Deciduous</td>
<td>Temperate (Spring-Frost Tolerant)</td>
<td>Temperate (Spring-Frost Tolerant) Cold-Deciduous Malacophyll Broad-Leaved Tree</td>
</tr>
<tr>
<td>Rosaceae</td>
<td>Shrub</td>
<td>Broad-Leaved</td>
<td>Cold-Deciduous</td>
<td>Temperate (Spring-Frost Tolerant)</td>
<td>Temperate (Spring-Frost Tolerant) Cold-Deciduous Malacophyll Broad-Leaved Tree</td>
</tr>
<tr>
<td>Thalictrum</td>
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<td>Temerarate</td>
<td>Temperate Drought-Intolerant Herb</td>
</tr>
<tr>
<td>Artemisia</td>
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<td>Small-Leaved</td>
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<td>Drought-Tolerant</td>
<td>Warm-Temperate Drought-Tolerant Shrub</td>
</tr>
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<td>Quercus</td>
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<td>Broad-Leaved</td>
<td>Evergreen</td>
<td>Warm-Temperate</td>
<td>Warm-Temperate Evergreen Malacophyll Broad-Leaved Tree</td>
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<td>Taxodiaceae</td>
<td>Tree</td>
<td>Needle-Leaved</td>
<td>Evergreen</td>
<td>Warm-Temperate</td>
<td>Warm-Temperate Evergreen Needle-Leaved Tree</td>
</tr>
<tr>
<td>Cyclobalanopsis</td>
<td>Tree</td>
<td>Broad-Leaved</td>
<td>Evergreen</td>
<td>Warm-Temperate</td>
<td>Warm-Temperate Evergreen Sclerophyll Broad-Leaved Tree</td>
</tr>
<tr>
<td>Lithocarpus</td>
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<td>Broad-Leaved</td>
<td>Evergreen</td>
<td>Warm-Temperate</td>
<td>Warm-Temperate Evergreen Sclerophyll Broad-Leaved Tree</td>
</tr>
<tr>
<td>Euphorbiaceae</td>
<td>Shrub</td>
<td>Small-Leaved</td>
<td>Evergreen</td>
<td>Warm-Temperate</td>
<td>Warm-Temperate Evergreen Sclerophyll Shrub</td>
</tr>
</tbody>
</table>
Table 3.9. Assignment of each PFT to a biome which is most likely to have produced that assemblage following Ni et al. (2010)

<table>
<thead>
<tr>
<th>Plant Taxa/Family</th>
<th>PFT Group Number</th>
<th>Code</th>
<th>Biome</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pinus</td>
<td>1</td>
<td>c-te.e.n.t</td>
<td>Cool-temperate evergreen forest</td>
</tr>
<tr>
<td>Tsuga</td>
<td>1</td>
<td>c-te.e.n.t</td>
<td>Cool-temperate evergreen forest</td>
</tr>
<tr>
<td>Poaceae</td>
<td>2</td>
<td>g</td>
<td>n/a</td>
</tr>
<tr>
<td>Cyperaceae</td>
<td>3</td>
<td>s</td>
<td>n/a</td>
</tr>
<tr>
<td>Carpinus</td>
<td>4</td>
<td>te-fa.cd.mb.t</td>
<td>Temperate Deciduous broadleaf forest</td>
</tr>
<tr>
<td>Betula</td>
<td>5</td>
<td>te-ft.cd.mb.t</td>
<td>Temperate evergreen broadleaf and mixed forest</td>
</tr>
<tr>
<td>Rosaceae</td>
<td>5</td>
<td>te-ft.cd.mb.t</td>
<td>Temperate evergreen broadleaf and mixed forest</td>
</tr>
<tr>
<td>Thalictrum</td>
<td>6</td>
<td>te-dt.fb</td>
<td>Temperate Grassland</td>
</tr>
<tr>
<td>Artemisia</td>
<td>7</td>
<td>wt-di.fb</td>
<td>Warm temperate drought-tolerant shrub/forb</td>
</tr>
<tr>
<td>Quercus</td>
<td>8</td>
<td>w.t.e.mb.t</td>
<td>Warm temperate evergreen broadleaf and mixed forest</td>
</tr>
<tr>
<td>Taxodiaceae</td>
<td>8</td>
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<tr>
<td>Cyclobalanopsis</td>
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<td>Lithocarpus</td>
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<td>Euphorbiaceae</td>
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<td>w.t.d.n.t</td>
<td>Warm temperate evergreen broadleaf and mixed forest</td>
</tr>
<tr>
<td>Ericaceae</td>
<td>9</td>
<td>di.sl.lhs</td>
<td>Temperate Grassland</td>
</tr>
</tbody>
</table>
Figure 3.25 MG1 percentage pollen summary of PFT classification as classified in Table 3.9 above.
Figure 3.26 MG1 pollen concentration summary of PFT classification as classified in Table 3.9. PFTs 1 and 5 are quoted as $10^3$, while PFTs 2, 3, 4, 7 and 8 are quoted as $10^2$. 
Zone P-1 (12.6-11.9 kcal.yr BP; 383-322 cm) is characterised by an increasing trend in dominance of temperate evergreen broadleaf and mixed forest taxa. These account for 40-50% of the pollen record during this period, with average pollen concentrations of approximately 10,000-15,000 grains/g. Other notable contributors to the record are PFTs 1, 3 and 4 which combined, account for almost 35% of the record. An abrupt drop in PFT5 dominance (due to the decline in Betula pollen) is seen at approximately 12.0 kcal.yr BP. PFT 1 i.e. cool-temperate evergreen forest taxa become the primary contributors to the MG pollen record reaching highs of 55% and almost 30,000 grains/g. Notable peaks in the upper portion of this zone can be seen in PFT 8, indicating a slight increase in the abundance of evergreen broadleaf and mixed forest taxa in the catchment.

Zone P-2 (11.9-8.3 kcal.yr BP; 322-232 cm) sees a recovery in PFT 5, reaching highs of >60% which is subsequently followed by a general declining trend throughout the remainder of this zone. PFT 5 remains the primary contributor to the pollen record during this period with pollen concentrations ranging from 10,000-25,000 grains/g, indicating a return to temperate climate conditions. However, the increased levels of PFT 1 from approximately 9.4-8.3 kcal.yr BP (270-232 cm) suggest a shift to cooler conditions as indicated by an increase in evergreen species (Pinus and Tsuga).

Zone P-3 (8.3-5.8 kcal.yr BP; 232-172 cm) is characterised by a further decline in PFT 5 taxa which fluctuate between 25-35%. Fluctuations in PFT1 pollen are also evident, ranging from approximately 20-45%. Increased contributions from PFT 8 indicate an increased presence of warm evergreen broadleaf and mixed forest taxa and favourable conditions for these taxa in the catchment during this period.

Zone P-4 (5.8-2.3 kcal.yr BP; 172-85 cm) sees a notable peak in cool-evergreen forest taxa (PFT1) at approximately 3.4 kcal.yr BP reaching a high of almost 60% with pollen concentrations standing at approximately 20,000 grains/g. A notable increase in temperate deciduous broadleaf forest taxa appears in the record from approximately 5.2-3.4 kcal.yr BP (160-120 cm). Increased contributions also come from PFT 3 (Cyperaceae pollen) along with PFT 8 (warm temperate evergreen broadleaf and mixed forest). PFT 5 values reach their lowest of the entire MG1 record.

In terms of pollen percentage, PFT 8 (Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus & Euphorbiaceae pollen) begin to dominate the record in Zone P-5 (2.3-0 kcal.yr BP; 85-0 cm) following a decline in PFT 1 pollen levels (Pinus & Tsuga). In terms of pollen concentration contributions to the pollen record, a significant proportion also comes from PFTs 2 (Poaceae), 3 (Cyperaceae) and 7 (Artemisia) while contributions from PFT 5 remain low. A general declining trend in PFT 2 pollen is noted however, from approximately 1.7 kcal.yr BP.

3.5 Cosmogenic $^{10}$Be analysis

3.5.1 $^{10}$Be concentrations

The $^{10}$Be concentrations measured in bedrock, fluvial sediment samples and lacustrine palaeo-deposits of quartz are presented here. In general, high $^{10}$Be concentrations indicate low erosion rates and vice versa. The results of $^{10}$Be concentrations in the majority of the palaeo-deposits reveal relatively consistent concentrations ranging from 0.17-0.2 atoms g$^{-1}$ (Table 3.11). However, sample MG3c indicates a
concentration of one order of magnitude variation at approximately 3.8 kcal.yr BP with a much lower $^{10}$Be concentration of 0.07 atoms/g $^{-1}$. The modern $^{10}$Be concentrations measured from the bedrock sample (MGCR), fluvial site A (MGDR) and fluvial site B (MGMR) (see Figure 2.16) were 0.35, 0.22 and 0.10 atoms/g $^{-1}$ respectively (Table 3.11, Table 3.12). As outlined in section 2.6.6, the methods and terms described in Balco et al. (2009) were then followed for the steady erosion-rate calculation methods based on these $^{10}$Be concentrations using the CRONUS-Earth online exposure age calculator, Version 2.2 (Balco et al., 2008). The results of these calculations are presented in the following section.
Table 3.11 \(^{10}\)Be concentrations measured from in pure quartz separates (lake palaeo-sediments, in-situ bedrock erosion rates & 0.25-0.5mm sizes sediment)

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Mid-point Sample Age kcal.yr BP (Based on Unified Core Chronology)</th>
<th>Mid-point Sample Depth cm (Based on Unified Core Chronology)</th>
<th>(^{10})Be/(^{9})Be ABS ratio (1E(^{-15}))</th>
<th>Error (abs)</th>
<th>Error (%)</th>
<th>Sample Mass (g)</th>
<th>Carrier Mass (mg)</th>
<th>(^{10})Be Concentration (atoms/g.Q) (1E(^{6}))</th>
<th>(^{10})Be Concentration (atoms/g.Q) (1E(^{6})) Error</th>
</tr>
</thead>
<tbody>
<tr>
<td>MG3a</td>
<td>3.0</td>
<td>112</td>
<td>260.52</td>
<td>5.85</td>
<td>2.25</td>
<td>38.88</td>
<td>0.39</td>
<td>0.18</td>
<td>0.006</td>
</tr>
<tr>
<td>MG3b</td>
<td>3.4</td>
<td>120</td>
<td>172.61</td>
<td>7.91</td>
<td>4.58</td>
<td>26.30</td>
<td>0.39</td>
<td>0.17</td>
<td>0.009</td>
</tr>
<tr>
<td>MG3c</td>
<td>3.8</td>
<td>109</td>
<td>178.38</td>
<td>7.07</td>
<td>3.96</td>
<td>65.28</td>
<td>0.39</td>
<td>0.07</td>
<td>0.003</td>
</tr>
<tr>
<td>MG4a</td>
<td>0.7</td>
<td>21</td>
<td>148.44</td>
<td>6.11</td>
<td>4.12</td>
<td>19.26</td>
<td>0.39</td>
<td>0.20</td>
<td>0.009</td>
</tr>
<tr>
<td>MG4b</td>
<td>1.2</td>
<td>35</td>
<td>579.72</td>
<td>11.70</td>
<td>2.02</td>
<td>80.54</td>
<td>0.39</td>
<td>0.19</td>
<td>0.006</td>
</tr>
<tr>
<td>MGCR (Bedrock)</td>
<td>Modern</td>
<td>Modern</td>
<td>668.01</td>
<td>111.42</td>
<td>16.7</td>
<td>50.251</td>
<td>0.39</td>
<td>0.35</td>
<td>0.06</td>
</tr>
<tr>
<td>MGDR</td>
<td>Modern</td>
<td>Modern</td>
<td>177.14</td>
<td>10.95</td>
<td>6.2</td>
<td>21.28</td>
<td>0.39</td>
<td>0.22</td>
<td>0.01</td>
</tr>
<tr>
<td>MGMR</td>
<td>Modern</td>
<td>Modern</td>
<td>296.89</td>
<td>27.87</td>
<td>9.4</td>
<td>80.21</td>
<td>0.39</td>
<td>0.01</td>
<td>0.01</td>
</tr>
</tbody>
</table>
3.5.2 Modern erosion rates in the Lake Mugecuo catchment

Steady erosion rates based on the \(^{10}\)Be concentrations measured in bedrock and fluvial sediment samples are presented here (Table 3.12). These data are reported with 1-sigma internal uncertainties which is the conventional format for \(^{10}\)Be rates (Balco, 2008). The modern catchment erosion rates calculated from fluvial sampling site A along with the bedrock sample revealed steady erosion rates of approximately 0.17 ± 0.02 mm/yr and 0.04 ± 0.01 mm/yr respectively. A notable difference was found in the modern catchment rate calculated from fluvial sampling site B which gave steady erosion rates of 0.32 ± 0.04 mm/yr.

Table 3.12 In-situ bedrock erosion rates (MGCR) and mean catchment erosion rates calculated from 0.25-0.5mm sizes sediment (MGDR & MGMR).

<table>
<thead>
<tr>
<th>Sample name</th>
<th>Latitude (DD)</th>
<th>Longitude (DD)</th>
<th>Atmospheric Pressure or Elevation (cm)</th>
<th>Thickness (cm)</th>
<th>Density (gcm(^{-2}))</th>
<th>Shielding correction</th>
<th>Mean Erosion Rate (mm/yr)</th>
<th>Erosion Rate Error Total (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MGDR</td>
<td>30.159</td>
<td>101.54</td>
<td>600.21</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>0.17</td>
<td>0.02</td>
</tr>
<tr>
<td>MGMR</td>
<td>30.155</td>
<td>101.53</td>
<td>632.38</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>0.32</td>
<td>0.04</td>
</tr>
<tr>
<td>MGCR</td>
<td>30.155</td>
<td>101.55</td>
<td>3831</td>
<td>0.96</td>
<td>2.6</td>
<td>0.45</td>
<td>0.04</td>
<td>0.01</td>
</tr>
</tbody>
</table>

3.5.3 Palaeo-erosion rates in the Lake Mugecuo catchment

Steady erosion rates for the Lake Mugecuo catchment over the last 3,800 years based on the \(^{10}\)Be concentrations measured in lacustrine palaeo-deposits of quartz are presented in Table 3.13 and outlined in Figure 3.27. These data are also reported with 1-sigma internal uncertainties. The younger of the sediments measured (MG4a & MG4b) were found to have similar steady erosion rates of 0.19 ± 0.02 mm/yr (MG4a) and 0.20 ± 0.02 mm/yr (MG4b). Similarly, the oldest sediments measured (MG3a & MG3b) were shown to have equal steady erosion rates of 0.21 ± 0.02 mm/yr and 0.22 ± 0.02 mm/yr, respectively. In contrast, a notable difference in palaeo-erosion rates is seen at approximately 3.8 kcal.yr BP where sample MG3c reveals an erosion rate of approximately 0.52 ± 0.05 mm/yr. This significant change in steady erosion rates likely reflects a perturbation in catchment geomorphic processes during this period and is further analyzed and discussed in Chapters 5 & 6.
Table 3.13. Palaeo-erosion rates in the Lake Mugecuo catchment

<table>
<thead>
<tr>
<th>Sample</th>
<th>Latitude (DD)</th>
<th>Longitude (DD)</th>
<th>Atmospheric Pressure or Elevation</th>
<th>Thickness (cm)</th>
<th>Density (g cm(^{-2}))</th>
<th>Shielding correction</th>
<th>Maximum Erosion Rate (mm/yr)</th>
<th>Erosion Rate Error Total (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MG4a</td>
<td>30.157</td>
<td>101.5</td>
<td>600.21</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>0.19</td>
<td>0.02</td>
</tr>
<tr>
<td>MG4b</td>
<td>30.157</td>
<td>101.5</td>
<td>600.21</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>0.20</td>
<td>0.02</td>
</tr>
<tr>
<td>MG3a</td>
<td>30.156</td>
<td>101.5</td>
<td>600.21</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>0.21</td>
<td>0.02</td>
</tr>
<tr>
<td>MG3b</td>
<td>30.156</td>
<td>101.5</td>
<td>600.21</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>0.22</td>
<td>0.02</td>
</tr>
<tr>
<td>MG3c</td>
<td>30.156</td>
<td>101.5</td>
<td>600.21</td>
<td>0.01</td>
<td>2.6</td>
<td>0.99</td>
<td>0.52</td>
<td>0.05</td>
</tr>
</tbody>
</table>

Figure 3.27. Unified core chronology with measured catchment-wide palaeo-erosion rates for the Mugecuo catchment over the last 3,400 years.
3.6 Reference climate record

The D4 δ¹⁸O record was interpolated to an evenly spaced data set and then transformed using a very low degree of smoothing using the filter function in Matlab v. R2011 in order to act as reference regional climate record against which the pollen and lake sediment data can be compared through time. The analysis of the climate, pollen and lake sediment data are discussed in Chapters, 4, 5 & 6. The original D4 δ¹⁸O record is presented in Figure 3.28 for reference and the smoothed D4 δ¹⁸O record utilised in this study is presented in Figure 3.29.

![Figure 3.28](image1.png)

**Figure 3.28.** Raw δ¹⁸O speleothem data taken from stalagmite D4 at Dongge Cave (25°17′N, 108°5′E) showing the shift in EAM intensity through the Holocene.

![Figure 3.29](image2.png)

**Figure 3.29** Smoothed D4 δ¹⁸O speleothem data taken from stalagmite D4 at Dongge Cave (25°17′N, 108°5′E) which is utilised as a regional Holocene climate record for this region.
Chapter 4: The MG1 pollen record

The MG1 Lake Mugecuo record is a sequence which provides an opportunity to explore the dynamics between changes in the East Asian Monsoon and palaeoecological changes following a transition from glacial to interglacial conditions (c. 12.6-0.0 kcal.yr BP) in Sichuan Province, Southwest China. It provides a complete palaeoclimatic record for this period, extending our knowledge and understanding of millennial-centennial scale climate and vegetation during the Holocene. This chapter aims to address the following research aims initially outlined in section 1.4:

- To identify the key environmental and climatic shifts that have occurred in the Lake Mugecuo area during the Holocene
- To examine the influence of these climatic shifts on plant functional diversity in the Mugecuo catchment i.e. what do these changes tell us about the plant community dynamics at Lake Mugecuo in general
- To compare the findings of these data with other regional records in order to provide a deeper insight into climate dynamics in Southwestern China

Information on the floral composition of the vegetation types and their relationships with environmental gradients (e.g. Hou, 1983; Walker, 1986; Xiwen & Walker, 1986; Prentice & Webb, 1998; Yu et al., 1998; Jiang et al., 1999; Chapman & Wang, 2002; Ni, 2003; Wang & Ni, 2005) at the Lake Mugecuo catchment was used for the interpretation of the fossil pollen spectra in order to answer these research questions. In section 4.1 each of the pollen zones are deconstructed and qualitatively analysed. Section 4.2 analyses the changes in plant community dynamics in the Lake Mugecuo catchment and critically assesses the method of grouping plant taxa into plant functional types to determine the effects of climate change on plant community dynamics. In section 4.3 the MG-1 record is compared to other regional proxy records to determine whether it supports the existing climate record for this region in order to assess its validity.

The potential relationship between changes in lake catchment dynamics i.e. erosion/deposition over time (based on the reconstruction of lake catchment dynamics which follow in Chapter 5) to the changes in catchment vegetation outlined in this chapter are further discussed and examined in Chapter 6. This comparison is made in order to disentangle the relationship (and/or influence) between changes that have occurred within the catchment based on lake proxies (i.e. erosion events, sedimentation rates, lake level change, river meandering), to the changes outlined in the pollen record and plant community dynamics during the Holocene.
4.1 Pollen interpretation

The dispersal distance of pollen grains is dependent upon a number of factors. These include; the productivity and dispersal mechanisms of the plant contributing to source of pollen; the fall speeds of the pollen grains being released; the grain morphology and size; the atmospheric conditions prevailing at the time of spore release and any barriers to movement i.e. mountains (Poska et al. 2011). As shown by theoretical models of pollen transport, the proportion of extra-regional pollen deposited in lake sediments increases with lake size (Jackson, 1994; Sugita, 1994; Jackson & Lyford, 1999). The long-distance component should be large when the regional pollen production is low, but it should be lower when the regional vegetation is dense and its pollen production is high (Herzschuh 2006a). In mountain regions, there is a tendency for samples to be allocated to forest biomes which occur at lower elevations than the sample site (Ni et al., 2010). Due to the nature of the topography of the catchment, it is likely that a significant proportion of the non-arboreal pollen component of the record entered the lake via wind transport (c.f. Sugita, 1994; Davis, 2000). The majority of the taxa in the record utilise anemophilous transport (e.g. Pinus, Quercus, Betula, Carpinus, Taxodiaceae, Poaceae, Cyperaceae) and thus it is likely that the main source of pollen to the lake was via wind transport. These issues associated with pollen interpretation are further discussed in the sections below.

4.2 The MG1 pollen record: Zones P-1 to P-5

Based on the ecological preferences of major taxa (Hou, 1983; Walker, 1986; Xiwen & Walker, 1986; Prentice & Webb, 1998; Yu et al., 1998; Jiang et al., 1999; Chapman & Wang, 2002; Ni, 2003; Wang & Ni, 2005), the palynological record of the MG1 core sediments spanning the last 12,600 years reveals a detailed history of vegetation and climate variation in this region during the Holocene as outlined in the following sub-sections. The smoothed D4 δ¹⁸O record is also discussed here in relation to the vegetation changes at Lake Mugecuo during the Holocene.
Figure 4.1 MG1 Pollen percentages of arboreal (AP) and non-arboreal (NAP) taxa including trees, shrubs, herbs, grasses and sedges. Spores/ferns and aquatics are also included. Due to the large number of species identified, taxa with an abundance <0.1% are excluded from this diagram. Pollen zones (derived using CONISS including taxa with abundances of >0.1%), core stratigraphy and calibrated AMS $^{14}$C radiocarbon dates are also included for reference along with the smoothed D4 δ$^{18}$O speleothem data taken from stalagmite D4 at Dongge Cave (25°17′N, 108°5′E) which is utilised as a regional Holocene climate record for this region.
4.2.1 Zone P-1 (12.6-11.9 kcal.yr BP; 383-322cm)

Temperate cold-deciduous broadleaved taxa (PFT-5) dominate this period of the Lake Mugecuo catchment vegetation history. The dominance of *Betula* (a cool-wet indicator, Hou 1983) and increased abundances of deciduous oak pollen (*Quercus*) signifies a relatively cold and wet winter climate during this period. *Betula* is a broad-leaved deciduous taxa, generally found in temperate regions as it is spring-frost tolerant (Hou, 1983). It is an alpine pioneer species which can tolerate minimum temperatures of -40°C (Woodward, 1996), and is generally associated with stage 1 forest succession patterns as it signifies stand re-initiation by colonising grassland areas (Martin et al., 1996). The dominance of *Betula* during this period likely represents a recent influx of birch-dominated forest taxa to the catchment as Last Glacial Maximum (LGM) conditions weakened with the onset of the Holocene and more favourable climatic conditions (see section 4.4.1 for further discussion of LGM climate in Southwest China). High levels of *Betula* pollen represent characteristic elements of the montane forest belt on the south eastern Tibetan Plateau the understorey of which is mostly composed of woody Rosaceae species (Kramer et al., 2010). The PFT classification presented in Chapter 3 suggests a close affinity between these two taxa in the Lake Mugecuo catchment. *Betula* pollen generally reflects the real distribution of the taxa (Yu et al., 2001) i.e it is not over- or under-represented in the pollen archive. The *Artemisia/Cyperaceae* ratios also indicate a dominance in temperate conditions during this period. The record peaks dramatically ca. 12.2 kcal.yr BP (369cm). The high value (~26) of Art/Cyp ratios ca. 12.2 kcal.yr BP may be indicative of a brief period of increasingly drier conditions as the contributions of sclerophyllous *Cyclobalanopsis* and spring-frost avoiding *Carpinus*, are also a likely indication of relatively warm and dry summer conditions. *Thalictrum* is a taxon which generally tolerates mean annual temperatures of <1°C with a mean annual precipitation of 200-300mm (Hou, 1983). Low temperatures are overcome during sunny weather by intense insolation (Hou, 1983). The presence of *Thalictrum* in the record therefore also points to warm, dry summer conditions throughout the period 12.6-12.5 kcal.yr BP.

The Younger Dryas (YD) event is a millennial-scale cold period and dry (~1300 year duration) between approximately 12,900 and 11,500 yr BP, which punctuated the last glacial termination (Ma et al. 2012). Speleothem records (including the D4 δ¹⁸O speleothem data taken from Dongge Cave, Figure 4.1) show a period of weakened East Asian Summer Monsoon (EASM) during the Younger Dryas event (Wang et al., 2001, 2007; Yuan et al., 2004; Dykoski et al., 2005; Sinha et al., 2005; Shakun et al., 2007; Liu et al., 2008; Yang et al., 2010). A notable increase in warm-temperate evergreen broadleaf and mixed forest taxa occurs in the MG1 record during the sharp decline in *Betula* pollen at 12.3 kcal.yr BP. The combined traits of these contributing taxa (based on classifications by Hou, 1983) to the pollen record suggest that the dramatic drop in *Betula* pollen is likely a result of a dramatic shift in effective moisture conditions at Lake Mugecuo i.e. there is a rapid shift to much drier winter conditions which are characteristic of similar changes seen in other records in Southwest China in the Younger Dryas chronozone (e.g. Ma et al., 2012). The nature of the
Younger Dryas period in China is still very much open to debate and is further discussed in section 4.4.1.

Betula species are generally fast-growing and short-lived trees (40-50 years) (Forrest, 2006) which would explain a quick response time to major changes in climatic conditions. At high elevations, conifers are more sensitive to temperature variations during winter as well as during the summer season (Shao & Fan, 1999; Liang et al., 2008), which is a likely reason for the sudden peak in Pinus pollen of almost 50% during this period. This dramatic shift in the MG pollen record lasts approximately 300 years, providing an interesting example of a pronounced centennial-shift in the palaeoecological record in the Early Holocene. The sedimentation rates (see Figure 5.1) remain relatively stable in the core during this period indicating that this event is not the result of a dramatic change within the lake system itself.

The Shannon-Wiener index can vary from 0, for a community with one species only, to values of 4 or greater in species-rich forests. Pollen diversity during this period at Lake Mugecuo stands at an average of ~2 indicating a relatively diverse catchment vegetation cover which is likely due to increasingly favourable growth conditions with a weakening of LGM climatic conditions. However, the Shannon-Weiner diversity index (H') indicates a general declining trend in species diversity in this zone, but this is most likely attributed to the dominance of Betula pollen in the overall MG pollen spectrum as sedimentation levels remain relatively stable during this period. However, a slight increase at approximately 12.3 kcal.yr BP may be as a result of a change in the vegetation composition in the lake catchment following the Betula crash which encouraged competition amongst taxa to gain a foothold in the catchment and thus, an increase in species diversity in the pollen archive. Although it would appear that arboreal species dominate the record at this stage based on overall AP% and concentration, an AP/NAP ratio of ~2.5 indicates that a substantial stand of grassland vegetation is present in this early section of the MG1 record. A combined contribution of Poaceae and Cyperaceae pollen of approximately 15% also provides evidence to support this along with the highest contribution of Thalictrum pollen of the entire record.

The upper portion of this section of the record (~12.5 to 11.9 kcal.yr BP) is characterised by a gradual decline in temperate evergreen broadleaf and mixed forest species and an increase in cool-temperate evergreen forest taxa. The climate of the period was likely to have become increasingly cooler and with moderately humid summer rainfall conditions. Montane Pinus spp in the Hengduan Mountains are relatively thermophilous and resistant to drought growing vigorously and extensively at an altitude of 3,000-3,400m (Walker, 1986). The main Pinus spp which are found growing in Sichuan today (Pinus yunnanensis and Pinus armandi) can tolerate mean winter temperatures of ca. -3°C (Chinese Forestry Academy, 1982-1985), they are extremely tolerant of spring drought conditions and occupy regions where annual precipitation ranges from between 600-1500mm (Jarvis, 1993). Carpinus is a taxon that can tolerate mean annual temperatures between 1-8°C. Thus, increased abundances of Carpinus pollen are also an indication of cool conditions. Carpinus is a taxon usually indicative of transitional type of vegetation between evergreen conifer and deciduous broad-leaved forests.
suggesting a very mixed forest community was present in the catchment at that time. Interestingly, it can be seen to peak slightly throughout the entire MG1 record during transitional periods of Betula- and Pinus-dominated forest.

4.2.2 Zone P2 (11.9-8.3 kcal.yr BP; 322-232cm)

Betula makes a dramatic recovery in the early portion of this zone, once again dominating the pollen record suggesting a revert back to temperate conditions with wet winters. From 11.9 kcal kyr BP a significant increase of mainly Betula tree pollen likely indicates a strong upward shift of the treeline (c.f. Lake Naleng Kramer et al. 2010). The subsequent decline in Betula likely signifies a gradual change in climatic conditions over time and the onset of Holocene conditions in the lake Mugecuo record characterised by increasingly drier winter conditions and a weakening in the EASM (Figure 4.1). The D4 δ¹⁸O speleothem data indicates that the EASM is at its weakest intensity of the Holocene during this period (Dykoski et al., 2005). The increasing levels of Pinus pollen are a likely indication of secondary succession by Pinus following the decline of Betula during this period. The succession of Pinus implies that the climate is becoming increasingly colder and drier as Pinus is a hardy species with conical needles which enable the tree to survive in environments where water availability is low (Thomas et al., 2007). An increase in Art/Cyp ratios supports this as Artemisia is a cold and drought-tolerant non-arboreal taxa (Jarvis, 1993). It tolerates dry and cold winters and annual precipitation levels as low as 150-250mm (Hou, 1983).

Thus, it is inferred that over time the vegetation was changing from a temperate evergreen broadleaf and mixed forest seen previously, to a predominantly cool-temperate evergreen forest. Increased diversity within the pollen record is apparent from approximately 11.9-9.5 kcal.yr BP which corresponds to declining sedimentation rates (see Figure 5.1 and for further discussion of non-climatic factors and ecological change see section 5.1) within the lake which likely resulted in a greater influx of pollen into the lake basin as sedimentation was slowing down. AP/NAP values indicate that the catchment remains aligned to an open woodland/grassland composition.

4.2.3 Zone P3 (8.3-5.8 kcal.yr BP; 232-172cm)

A further decline in Betula signifies cold and dry climatic conditions prevailing in the lake catchment with Pinus continuing to be a major contributor to the pollen record. However, some apparent fluctuations in both temperature and humidity are also displayed, with evidence of consistent contributions (averaging approximately 20%) to the pollen record from warm-temperate evergreen and broadleaf mixed forest taxa. Increased contributions come from Taxodiaceae, a family of hardy conifers, and warm-loving, needle-leaved trees (Hou, 1983) as well as from Tsuga, a mesic conifer which requires an annual precipitation of ~1000-1500 mm, but tolerates cold temperatures (Academia Sinica, 1983-1985; Editorial Board of Sichuan’s Vegetation, 1980). Tsuga, a warm-wet indicator species (Xiaoqiang et al., 2003) is strongly associated with moist temperate habitats and is often used as an indicator species of a dominance of summer rainfall (Jarvis 1993). It is thus interpreted that this
was a period of cold, dry winters and warmer summers with increased summer rainfall i.e. the catchment vegetation was subject to distinct wet and dry seasons. The catchment remains aligned to an open woodland/grassland composition based on AP/NAP values, although non-arboreal contributions to the pollen record are on the increase.

4.2.4 **Zone P4 (5.8-2.3 kcal.yr BP; 172-85cm)**

This period of the lake Mugecuo record sees a dramatic rise in cool-temperate evergreen forest species. Increases in Cyperaceae pollen are also associated with cool conditions, as high abundances of Cyperaceae are often associated with cool alpine meadow and shrubland in this region (Yu et al., 2000a). A decline in AP/NAP values signifies an increase in grassland/shrubland species in the catchment and a contraction of the forest component.

Relatively high contributions from warm-temperate evergreen broadleaf taxa are seen from approximately 5.2-3.4 kcal.yr BP prior to the sudden increase in *Pinus* pollen at approximately 3.6 kcal.yr BP. A climatic shift to a cooler and/or drier climate at this point would be extremely beneficial to *Pinus*, providing it with a competitive edge over other more temperature-sensitive taxa (Shen et al., 2005; 2006), albeit with a reduced forest cover based on rising NAP values. Fluctuations in *Tsuga* following this shift reveal that summer rainfall continues to dominate, although temperatures are cooler than previously. A dramatic decline in pollen diversity occurs at 120cm but TPC levels remain relatively stable. This suggests that the sudden influx of *Pinus* pollen is casting an overriding signal in the pollen spectrum whereby it is lowering the diversity levels of pollen being preserved in the lake during this period due to the sheer volume of pollen being dispersed close to the lake basin.

4.2.5 **Zone P5 (2.3-0 kcal.yr BP; 85-0cm)**

A declining trend in cool-temperate pollen abundances characterise this period in the uppermost section of the MG1 core, with warm-temperate taxa gaining a strong foot-hold in the catchment record. A decrease in evergreen needle-leaved species in this zone is inferred to reflect a decrease in precipitation during the growing season. Sclerophyllous taxa of Southwestern Sichuan require a moderate winter climate but can withstand the drought stress conditions of the spring and early summer of Southwestern China (Jarvis, 1993). An increase in these broadleaved evergreen sclerophyllous taxa (PFT-8), provide evidence of a continuous decrease in late-spring and summer monsoon precipitation and a rise in winter temperatures (as per Jarvis, 1993). Increased abundances of Euphorbiaceae pollen from 2.3 to 1.4 kcal.yr BP are also interpreted as an indication of increased drought stress during the winter (Shen et al, 2006). Poaceae species tend to grow in arid valleys and alluvial-fluvial piedmont plains in southern Tibet (Li et al., 1985) and are considered eurytopic species (Huang, 2000). This type of grass vegetation in generally found in the temperate zone under a semi-humid climate with an annual precipitation of 350 to 500mm (Jarvis, 1993). Increased abundances of Poaceae pollen in the MG1 record during this period therefore also point to a warming climate in the Late Holocene.
Increased diversity within the record at this time is likely due to the combination of a) a decline in *Pinus* pollen which is often a dominant signal in any pollen profile as it is a prolific pollen producer and b) increasing sedimentation rates (Figure 5.1) which may have resulted in the increased influx of arboreal pollen into the lake via fluvial transport. A further decline in AP/NAP ratios indicate that the catchment is predominantly composed of grassland vegetation. However, based on these values, the increased pollen diversity values and the modern catchment vegetation stands, it is more likely that the true pollen signal is being distorted in terms of how pollen is transported to and archived within the lake as there still remains a large arboreal pollen component present in the pollen record. *Pinus* and *Quercus* often seen are known to be over represented in pollen spectra (Jarvis and Clay-Poole, 1992; Yu et al., 2004) and, in addition *Pinus* pollen, may be regarded as a sign for long-distance transport. However, *Quercus* values remain comparatively low throughout the Holocene whilst it is also difficult to argue that *Pinus* is over-represented in the Mugecuo record based on the changes of alignment of the vegetation which have most likely been in response to climatic change. It should be noted at this point however, that based on the dominance of arboreal pollen throughout the Holocene record and the composition of the modern catchment, it is highly unlikely that over time drastic changes in the vegetation stands would have occurred. A major change in the catchment vegetation distribution is highly limited due to the slope and aspect of the catchment valley which limits the availability of light and nutrients to vegetation. Therefore it is likely that a forest component and alpine meadow component have co-existed in the Mugecuo catchment since beginning of the MG1 record.

4.3 Palaeo-vegetation community dynamics in the Lake Mugecuo catchment

Changes in plant community dynamics in the lake Mugecuo catchment are discussed here on the basis of the PFT classification outlined in section 3.4.6 and the PCA ordination results from section 3.4.3.

4.3.1 A Reconstruction Of The Plant Community Dynamics In The Mugecuo Catchment

The changes in plant community composition are studied here on the basis that certain plants can be grouped together based on their association to certain environmental conditions. Based on this, the primary biomes recognised in the MG1 Holocene pollen record are cool-temperate evergreen forest, temperate mixed (evergreen and broadleaf) forest and warm-temperate mixed (evergreen and broadleaf) forest. An overview of the ecological tolerances of these biomes is outlined in Table 4.1.
Table 4.1 Basic concepts of global bioclimatic zonation adapted from Box & Fujiwara (2012).

<table>
<thead>
<tr>
<th>Zone</th>
<th>Basic Concept</th>
<th>Temperature Limits</th>
<th>Significance &amp; Vegetation Response</th>
</tr>
</thead>
<tbody>
<tr>
<td>Warm-Temperate</td>
<td>Mild winters with some frost</td>
<td>Light-moderate frost most years, to about -15°C; winter means &gt; 0°C</td>
<td>Vegetation still well developed and evergreen (more or mostly deciduous if colder or drier)</td>
</tr>
<tr>
<td>Temperate</td>
<td>Frost significant, summers not too hot</td>
<td>Frost every year, to &lt; -15°C; winter means &gt; or &lt; 0°C</td>
<td>Vegetation mainly winter-deciduous (conifers secondary or disclimax)</td>
</tr>
<tr>
<td>Cool-Temperate</td>
<td>Temperate but with cool summer</td>
<td>Temperate but with summer means &lt; about 20°C winters milder if oceanic</td>
<td>Vegetation mainly deciduous + conifers, with broadleaf evergreens if winter milder (as in Southern Hemisphere)</td>
</tr>
</tbody>
</table>

Overall, the Holocene pollen record of Lake Mugecuo plant community dynamics is generally dominated by tree taxa including *Betula, Pinus, Quercus, Taxodiaceae, Cyclobalanopsis* and *Lithocarpus*. To gain further insight into vegetation changes over time and to confirm the qualitative designation of the various taxa into specific groups, the pollen types representative of the plant functional types (PFTs) and biomes as per Ni et al. (2010) were subject to PCA analysis which gives an indication of the inter-species correlations within the pollen data set (Figure 3.21). Overall, three assemblages can be distinguished from the results of the PCA analysis: cool temperate (*Pinus*) evergreen forest (top left quadrant), warm temperate evergreen broadleaf and mixed forest (bottom left quadrant) and temperate evergreen broadleaf and mixed forest (*Betula*) (based on the biomization procedure used in Ni et al., 2010). These results suggest that there is an underlying gradient which explains the composition of the dataset. Most Holocene continental climate archives in Asia are lake sediments which permit the reconstruction of vegetation dynamics; climate changes are therefore reflected as changes in the effective moisture, which represents a combined signal of temperature and precipitation changes (Herzschuh, 2005). Based on the ecological affinities of the taxa used in the PCA analysis (Hou et al., 1983; Walker, 1986; Li & Walker, 1986), Axis 1 is interpreted to represent effective moisture, while Axis 2 represents changes in summer temperature. The first principal component (Axis 1) broadly separates malacophyllous taxa (e.g. *Betula, Carpinus, Quercus*) and drought intolerant taxa (e.g. *Thalictrum*) to evergreen taxa with sclerophyllous leaves (e.g. *Cyclobalanopsis, Pinus, Tsuga, Taxodiaceae*) along with drought-tolerant non-arboreal taxa such as *Artemisia* and *Euphorbiaceae*. The second principal component (Axis 2) broadly separates cold-tolerant taxa (for example evergreen needle-leaved *Pinus* and spring-frost tolerant Rosaceae) from sclerophyllous taxa preferring milder/warmer climatic conditions e.g. *Cyclobalanopsis, Lithocarpus, Taxodiaceae, Poaceae*. The first principal component captured 32% of the total variance in the
dataset while the second principal component captured 18% of the total variance, suggesting that effective moisture followed by changes in summer temperature are likely to have been the primary environmental variables driving vegetation change at Lake Mugecuo during the Holocene.

Sharp angles (less than 90°) between species arrows indicate strong positive correlations. The angle and direction of correlation indicated by each arrow broadly correlates to the PFT groupings within the pollen dataset (Figure 4.2). Exceptions to this are Betula, Quercus and Rosaceae in this instance. According to the classification outlined in section 3.8, Quercus is grouped with Taxodiaceae, Cyclobalanopsis, Lithocarpus and Euphorbiaceae as a warm-temperate mixed (evergreen and broadleaf) forest species. However, Quercus is described in the previous section as an indicator of cool and wet conditions which is consistent with the results of the PCA analysis where it falls into the upper right quadrant. Therefore, the classification of Quercus into a warm-temperate mixed (evergreen and broadleaf) forest PFT is not quite accurate for this particular study. Rosaceae and Betula according to the PFT classification should also be closely correlated and are characteristically found growing together on the Tibetan Plateau as previously stated. However, the results of the PCA analysis do not show a close correlation between these two taxa in the dataset. The data suggests that Rosaceae are more closely correlated with sclerophyllous, cold-tolerant taxa at the Lake Mugecuo catchment. It can be seen that increased levels of Rosaceae pollen appear in the MG1 record during colder and drier periods e.g. 7.1-4.7, 2.3-1.7 kcal.yr BP. The PCA analysis therefore highlights some initial weaknesses of the PFT classification method used in this study.

Figure 4.2 PCA species ordination plot for the MG1 pollen data set, Lake Mugecuo. Axis 1 separates sclerophyllous from malacophyllous taxa. Axis 2 separates cold-tolerant taxa from thermophilous taxa. A guide to plant taxa as they are classified into their respective PFTs are denoted by a dashed red circle.
The division of species in a PFT classification can often be divided into ‘structural’ and ‘interstitial’ species (Huston, 1994). The structural species have a major influence on the diversity of the interstitial species: in most cases the interstitial organisms would not be present in a given area without the structural organisms (Huston, 1994). An example of this in the Lake Mugecuo catchment is the association of *Betula* and understorey Rosaceae, and the association of warm-temperate mixed (evergreen and broadleaf) forest taxa (PFT 8) with warm-temperate sclerophyllous (rigid leaves) Euphorbiaceae shrubs which likely take the form of understorey vegetation when warm-temperate conditions prevail. Structural species, such as trees, provide the physical structure of the environment, which is used by the interstitial species and which are invariably smaller, such as epiphytes and understorey herbs (Duckworth *et al.*, 2000). The Lake Mugecuo record indicates that a forest component prevailed in the catchment over time which was also coupled with a relatively persistent alpine meadow component (*Poaceae, Thalictrum, Artemisia, Cyperaceae*) i.e. both tree and shrubs/grasses play an important functional role in the plant community dynamics of this catchment.

The last decade has seen extremely rapid growth of information on plant traits and their relationship to climate (e.g. Wright *et al.* 2004; Box & Fujiwara 2012). In terms of studying climatic change and the effects it has on functional diversity, predicting the favourable growth season is important because different inter-annual variability regimes may influence the balance of evergreen and deciduous plants (Ni *et al.* 2005). All plants require a period with temperatures sufficient for growth (Ni *et al.*, 2010). The seasonal distribution of precipitation can strongly affect the severity of drought experienced by the plants, and therefore the plant forms that are adapted to the climate (Neilson *et al.*, 1989). The distributions of phanerophyte (a perennial tree or shrub with dormant buds borne on aerial shoots; Raunkiaer, 1934) species are predominantly restricted by winter temperature extremes. Deciduousness in phanerophytes is widespread in climates where there is a clear division of seasons which are favourable or unfavourable to plant growth including regimes with mean temperatures <5°C in the winter months, and monsoonal regimes with pronounced wet and dry seasons (Harrison *et al.*, 2010). The dominance of PFT 5 (*Betula, Rosaceae*) in the lower section of the MG1 record (Younger Dryas to early Holocene; approximately 12.6-8.3 kcal.yr BP) indicates a clear division between seasons, characterised by a cold, wet winter climate with relatively warmer and drier summer conditions. In general, it can be said that deciduous plants predominate in climates with moderately long unfavourable periods and evergreen species in climates with year-round favourable growth conditions (Harrison *et al.*, 2010). Exceptions to this include temperate winter rainfall climates which favour evergreens, as the most effective strategy is to use soil moisture stored from winter to allow transpiration and photosynthesis in spring (Sala *et al.* 1997). There is a broad range of climates; temperature regimes in the case of cold-deciduous plants, and moisture regimes in the case of drought-deciduous plants, in which both evergreen and deciduous phanerophytes co-exist (Harrison *et al.*, 2010). Cool-temperate evergreen forest dominates the middle Holocene Lake Mugecuo record (approximately 8.0-2.3 kcal.yr BP) but notable contributions from temperate mixed (evergreen and
broadleaf) forest taxa prevail from approximately 8.0-4.8 kcal.yr BP. The relative success of the two strategies during this period is attributed to competition, the winning strategy (PFT 1; Pinus, Tsuga) maximizing the ratio of benefits to carbon and/or nutrient costs (c.f. Schulze 1982; Matyssek 1986; Kikuzawa 1996; Aerts & Chapin 2000; Givnish 2002). The decline in PFT 5 (Betula, Rosaceae) over time due to increased competition from PFT 1 (Pinus, Tsuga), closely followed by an increase in warm-temperate evergreen broadleaf and mixed forest indicates that changes in climatic conditions are occurring in the form of drier winters and warmer summers with increased summer rainfall i.e. it is likely that distinct wet and dry seasons still prevailed in the Mugucuo catchment at this time. Warm-temperate evergreen species dominate the Late-Holocene (4.8-0.0 kcal.yr BP) vegetation record at Lake Mugecuo and are indicative of a less seasonal division in the climatic regime with a rise in winter temperatures and decreased summer monsoon precipitation.

The terms “sclerophyll” and “malacophyll” (Schimper 1898, 1903) describe leaves that are rigid and soft, respectively. The relation of leaf density to climate is primarily linked to the mechanical properties of sclerophyll leaves which allow them to support transpiration longer during periods of drought (Oertli et al. 1990; Niinemets 2001; Lamont et al. 2002). A distinct increase in sclerophyllous taxa in the Late-Holocene at Lake Mugucuo marks the establishment of a strongly seasonal rainfall regime in the region with extended periods of drought. The advantage of sclerophyllous leaves is greatest in temperate climates with winter rain, where soil moisture deficits coincide with maximum potential productivity; and (b) the high construction costs of sclerophyllous leaves are inconsistent with a short leaf lifespan, so sclerophylls are absent wherever winters are too cold to support the broadleaved evergreen habit (Harrison et al., 2010) i.e. seen during periods indicative of Younger Dryas conditions and early Holocene conditions at Lake Mugucuo. Long-lived evergreen leaves are expensive to construct, but can be amortized over several growing seasons while short-lived deciduous leaves are inexpensive to construct, but must be constructed again each year (Harrison et al., 2010). The leaves of the temperate needle-leaved evergreens and the buds of temperate broad-leaved deciduous species can tolerate extremes of -20°C to -30°C, and in some cases down to -42°C. This gives an indication of how low the winter temperatures at Lake Mugucuo during the early to mid-Holocene may have been. The environmental selection for certain traits by species will often result in a trade-off of some form or another i.e. the traits that characterise a PFT can often convey both advantages and restrictions. Sclerophyll leaves can extract water from dry soils at low water potential and are typically long-lived, allowing their high construction costs to be amortized over a number of years (Harrison et al., 2010). However, their photosynthetic capacity (even on a per area basis) is low because of internal shading within the leaves and/or high mesophyll resistance to CO₂ diffusion (Lloyd et al. 1992; Niinemets 1999) i.e. this trade-off means that sclerophylls can dominate in dry environments, but they do not compete effectively in wetter environments. Chilling requirements represent a cost for cold tolerance because photosynthesis is shut down during dormancy, and because cold-tolerant species may be excluded from warmer environments (Niinemets 1999; Harrison et al., 2010). The presence of a woody stem will allow a tree to transport water efficiently from its roots to leaves, but increasing
amounts of carbon then have to be allocated to maintain living sapwood and fine roots in order to avoid cavitation in taller stems (Sperry et al. 2008). A trade-off such as this results in trees being competitively favoured in productive environments, but unviable in less productive environments. The general point being raised here is that trade-offs are universal, and promote the diversity of plants across bioclimatic gradients (Harrison et al. 2010). The results of such trade-offs have resulted in the close association/relationship of select plant taxa with each other from the MG1 pollen record which allow us to explore the broad-scale changes in the relationship between the East Asian Monsoon and changes in plant community dynamics during the Holocene. The pollen spectrum and association of certain plant taxa indicate that the Mugucuo climate is and has been a relatively productive environment for vegetation since at least 12.6 kcal.yr BP with a relatively rich array of plant diversity.

4.3.2 The link between pollen records and plant functional types- a viable concept?

There is a recognized relationship between PFTs and climate, particularly at the global or biome scale (e.g., Woodward, 1987; Prentice et al., 1992; Box, 1981; 1996). However, there is still very limited knowledge about how different plant traits and different combinations of environmental factors will interact in determining whole-plant responses and thus responses to climate change at a higher scale of organization (Díaz and Cabido, 1997) An approach based on PFTs simplifies the system and enables the consideration of the effects and interactions between species at different levels of organization (Duckworth et al., 2000). The method of grouping pollen taxa/species/families into plant functional types ends with an assignment of that PFT to a biome which is most likely to have produced that assemblage i.e. the prevailing climatic conditions can be inferred from the dominant plant fossil assemblage found in the lake sediment archive at a given time slice. However, pollen-productivity and hence its representation in the pollen record depends on the plant species (Dallmeyer et al., 2011). The phenomenon of upward transport of pollen to higher elevations in mountain regions has been noted in the previous biomization of data from China (Yu et al., 2000) and was convincingly demonstrated as the major cause of mis-allocations of samples in Japan (Takahara et al., 2000). Most subalpine shrubs, for instance, are poor pollen-producers and thus have a low representation in pollen spectra (van Campo et al., 1996; Herzschuh and Birks, 2010). Hence, shrub coverage could be underestimated in the method of biome reconstructions utilised in this study which may be the reason that Rosaceae did not correlate with Betula in the PCA ordination analysis.

In general, most pollen can only be identified to genus or family level. Therefore, the assignment of pollen taxa to biomes within the biomisation techniques is a strong simplification of the natural conditions (Dallmeyer et al 2011). Species of a given plant functional type share similar morphological, physiological or life-history traits, with the differences between members within one functional type being smaller than those between types (Smith and Huston, 1989). Taxa such as Castanopsis, Cyclobalanopsis and Lithocarpus are characteristic of the warm-temperate broadleaf and mixed forests have leaves that are physically intermediate between malacophyll and sclerophyll (i.e. coriaceous leaves) (Ni et al., 2010). However, an intermediate category such as this does not
exist in the current PFT scheme outlined by Ni et al. (2010) and as a result these taxa are allocated to both a mixed malacophyll and sclerophyll PFT with a consequent loss of discriminatory power. In the end only the dominant tree PFTs and some of the more highly characteristic elements of the secondary PFTs were used to define forest biomes following methods by Prentice et al. (1992). However, the assignment of taxa to PFTs in this study is based on the analyses of the biomization of modern and palaeo-samples from China by Ni et al. (2010) which suggests that the current scheme is likely to yield reasonably good reconstruction of the broad-scale patterns of palaeo-vegetation over time.

Another important factor that needs to be taken into consideration when studying the effects of climatic change on plant community dynamics is that species migrations, as well as extinctions of individual plants (especially trees), take some time after any climatic change; therefore, the natural plant succession is relatively slow (Skrzypek et al., 2009). According to Delcourt and Delcourt (1987), in the case of trees, a change may take place within hundreds of years. However, of particular relevance here is the concept of redundancy – some species may substitute for each other in terms of ecosystem function (Walker, 1992). Because species exhibit unique responses to environmental change (e.g., Huntley, 1991), any change severe enough to cause extinction of one species is unlikely to eliminate all members from a functional group (Walker, 1992). If one species goes extinct, another from the same functional group can take its place and the loss of one species will have less effect if others from the functional group are retained (Lawton and Brown, 1993). A major claim of the PFT approach is the elimination of both noise and redundancy, thus simplifying plant community description (Duckworth et al 2000). Ecosystem functional types are a broader grouping, based on the manner in which plants modify their environment and hence the ecosystems in which they are found, by changing the temporal or spatial nature of the system response (Shugart, 1997). The more species there are in a functional group, the more stable the system is and the less serious the consequences of extinction are likely to be (Walker, 1992; Huston, 1994; Chapin et al., 1997). The use of an ordination package such as CANOCO (ter Braak, 1990) allows the user to downweight the importance of infrequently occurring species in the analysis or make them passive, acknowledging the inherent noise in species data. However, by reducing the number of entities in the PCA i.e. rarer species (in this case those with pollen percentages <0.5%), there may be a significant loss of information in terms of the changes in plant functional groupings over time as species migrate or go extinct. Such changes are overlooked in this case as the PFT classifications remain static through time in this study-regardless of the potential loss or gain of species. Although we gain a broad-scale picture of the major changes in dominant PFT groupings over time, we may be missing out on the smaller-scale changes in the true plant community dynamics of the catchment. On the other hand, the main advantage of using such a PFT classification scheme is that it is globally applicable and allows for a relatively straightforward comparison of biome reconstructions among regions and continents (Harrison et al., 2010; Ni et al., 2010). However, further study is required to address the potential trade-offs between looking at small and larger-scale dynamics.
A comparison of the Lake Mugecuo catchment pollen record to the existing climate record for the region

The MG-1 record is discussed here in relation to other regional proxy records of Holocene climate and vegetation change. Sub-sections are used to compare these proxy records in terms of the Late-Glacial/Early Holocene transition, mid-Holocene and late-Holocene changes.

4.4.1 The Late-Glacial/Early Holocene transition

The Late Glacial (approximately 20-10 ka ago) was a period of cold and dry conditions on the Tibetan Plateau (Herzschuh et al., 2009). The Younger Dryas (YD) event is a millennia-scale cold period (~1300 year duration) between approximately 12,900 and 11,700 yr BP, which punctuated the last glacial termination (Ma et al., 2012). One of the most important points to be made about the Younger Dryas is that it shares the characteristics of the 24 or so millennia-duration Dansgaard-Oeschger (DO) cold spells that punctuate the glacial portion of the Greenland ice-core record (Stuiver and Grootes, 2000; Mangerud et al., 2009). However, according to Liu et al., (2013, in press) the regional climate correlation within the Northern Hemisphere in the cold/dry mid-Younger Dryas event (YD) remains elusive. Modeling and geological records have shown that climate conditions in the mid-Younger Dryas (12.5–11.6 ka) were unstable (Manabe and Stouffer, 1997; Zhou et al., 1999, Wang et al., 2001 and Nakagawa et al., 2006). In the Lake Mugecuo record, evidence of Younger Dryas climatic conditions appear in the pollen record at approximately 12.3 kcal.yr BP with drier winter conditions prevailing which is a defining characteristic of Younger Dryas chronosequence records in Southwest China (c.f. Ma et al., 2012). Previously published high-resolution, precisely dated speleothem records and a peat record from southern China indicate that EASM precipitation increased shortly after the Younger Dryas (YD) at the onset of the Holocene, reached its highest level during the early Holocene, and then decreased gradually over the next several thousand years (e.g., Dykoski et al., 2005; Hong et al., 2005; Wang et al., 2005). Thompson et al (1989) place the Pleistocene-Holocene Transition at ~11,900 cal.yr BP on the Tibetan Plateau, while Jiang et al. (1999) suggested that widespread glacial conditions were not concluded until ~9 ka BP in this region. Evidence from the proxy record of Lake Qinghai indicates that on the north eastern Tibet-Qinghai Plateau the climate across the late-glacial/Holocene transition was unstable, showing abrupt shifts in hydro-climatic conditions on time scales of less than a millennium (Yu & Kelts, 2002). Other palaeoenvironmental records such as the Baxie loess-palaeosol (An et al., 1993) and the Guiyla ice cap (Thompson et al., 1997), indicate that short-term variations in climatic conditions on time scales similar to that at Lake Qinghai occurred at these sites, and that the onset of a warm and relatively wet early-Holocene climate was synchronous at about 10,000 ^14C yrs B.P (Yu & Kelts, 2002). The early Holocene intensification of the EASM resulted in a much wetter climate compared to Late-Glacial conditions (Xingqi et al., 2008). Evidence of the LGM-Early Holocene transition can be seen where Betula dominates the early part of the Lake Mugecuo MG1 record which is an indication that extremely cold temperatures and wet winters dominated the Lake Mugecuo catchment during this
period. According to Shen et al. (1996), the presence of Betula and deciduous oaks coupled with the low values for Tsuga, Castanopsis and Cyclobalanopsis is a characteristic Late-glacial assemblage in Yunnan, reflecting the dominance of cool climatic conditions and weak summer rainfall. A very similar climatic pattern is also recorded at Eryuan Lake in Yunnan Province (Walker, 1986; Li and Walker, 1986). Birches grew widely in Northeast China, the Inner Mongolian Plateau, the Shandong Peninsula, the Qinling Mountains and the eastern Qinghai-Tibetan Plateau ca. 10ka BP (Ren & Beug, 2002). Therefore it is argued here that the beginning of the MG1 record (ca.12.6 kcal.yr BP) covers (at least) part of the transitional period between Late-Glacial to Holocene climatic conditions with some evidence of Younger Dryas conditions during the decline of Betula pollen at approximately 12.3 kcal.yr BP.

At the beginning of the Holocene (~11.7 cal. ka), EAM precipitation increased dramatically, synchronous with an abrupt warming in the North Atlantic (Morrill et al., 2003). Studies of the speleothems from Dongge Cave (Yuan et al., 2004; Wang et al., 2005) and Shanhao Cave (Shao et al., 2006) suggest that the climate became quite humid at the beginning of the Holocene in Southwest China with a strong summer monsoon and weak winter monsoon (Yancheva et al., 2007). High precipitation during the early Holocene has also been inferred based on organic carbon isotopic records from Hongyuan peatland (Hong et al., 2003) and pollen records from lakes in modern marginal monsoon areas (Liu et al., 2002; Li et al., 2004). Zhao et al., (2009) using fossil pollen records from 31 sites in the monsoonal region of China showed that a humid climate generally characterized the early and middle Holocene. Several studies suggest that the Holocene EAM intensity variation generally follows insolation variations, with some thousands years of lag (Kutzbach, 1981; COHMAP members, 1988; Wang et al., 2005). The time lag has been attributed to the possible response of monsoon intensity to multiple forcing, including changes in precession-driven insolation, obliquity-driven insolation and ice volume (Clemens and Prell, 2007). Based on the pollen assemblage in the deepest section of the MG1 record, it is most likely that the onset of Holocene conditions occurred at approximately 11.9 kcal.yr BP when a decline in Betula and an increase in cool-temperate evergreen forest taxa indicate a climatic shift from cold, wet winters to increasingly cooler climatic conditions and increasingly drier winter conditions. According to Jarvis (1993b), the early Holocene was a period of increased summer temperatures, cold winters, and a more even distribution of rainfall throughout the year than subsequently in Southwest China. Contrary to this, the MG1 record indicates a more seasonal pattern of rainfall during the Early Holocene period characterised by drier winters and wetter summer conditions. Accordingly, Herzschuh et al. (2006a) proposed that high arboreal pollen counts in fossil spectra from Lake Zigetang on the central Tibetan Plateau were indicative of sparse vegetation around the lake, owing to cold and/or dry climatic conditions. A high abundance of arboreal pollen in the MG1 record during this time likely supports similar winter conditions at Lake Mugecuo. Forests composed predominantly of Betula are not seen today in the mountains of southwestern Sichuan, although they are found in the mountains of the north and north-west of the province where winters are much colder (-11°C) (Jarvis, 1993).
suggests that winter temperatures were somewhat similar in the early Holocene with Betula dominating the Lake Mugecuo vegetation. Winter temperatures likely increased gradually throughout the Holocene which resulted in the decline of Betula forest in the catchment to the lower presence of birch that is found in the modern Mugecuo record. Taking into account the modern ecology of the Lake Mugecuo catchment, it is most likely that Betula and the other arboreal pollen taxa dominant during this period occupied the north facing slopes and favourable sites with exposure patterns similar to today. Shrub/scrub and alpine meadow taxa likely dominated the flat, western valley floor of the catchment. Increased river activity during the early Holocene for Eastern Tibet was driven by high rainfall probably due to the re-establishment of the monsoonal regime, as well as high surface runoff due to sparse vegetation cover on surrounding slopes (Schlötz & Lehmkuhl, 2009). The increase of Betula in the Nianbaoeyeze Mountains in eastern Tibet during the early Holocene (10.6-9.8 kcal.yr BP) points to a successive establishment of deciduous forests on gradually consolidating soils (Schlötz & Lehmkuhl, 2009). A similar scenario appears to occur at Lake Mugecuo with Betula forest dominating from approximately 12.6-9.9 kcal.yr BP. The pioneer species of Betula in the Lake Mugecuo catchment likely benefitted from the seasonally high river discharges that were caused by the strong rainfall associated with the rapidly strengthening summer monsoon (c.f. Fan et al., 1996; Gasse et al., 1996; Hong et al., 2003).

In a study by Jarvis (1993a) pollen evidence from Lake Shayema Mianning County, was obtained to examine evidence of postglacial vegetation and climatic change in Southwestern Sichuan. The pollen record obtained from this lake suggests that there have been several changes in the vegetation in this region during the past 11,000 years. Approximately 11,000 yr cal.B.P., high percentages of Betula and deciduous oak pollen suggest that these cold- and drought-tolerant taxa dominated the landscape, while the occurrence of Tilia, Ulmus, Fraxinus and Carpinus pollen which are taxa that can tolerate mean annual temperatures of about 1° to 8°C, also act as an indicator of cool conditions (Jarvis 1993a). An increase in evergreen oaks and dry-tolerant species after 10 ka at Lake Erhai, Yunnan Province suggests a greater seasonality of rainfall and reflects a southward shift in the winter front across the region (Shen et al., 2006). At Lake Chen Co, Southern Tibet, relatively high humidity and cold temperatures predominated from 16.5-10.5 kcal.yr BP (Zhu et al., 2009) similar to the prevailing conditions at Lake Mugecuo during the same period. Zhu et al., (2009) report a shift from warm to cold temperatures and wet to drier conditions from 10.5-9.0 kcal.yr BP, which represents a similar shift in conditions that occurred earlier in the Lake Mugecuo catchment i.e. the dramatic decline in Betula pollen at 12.3 kcal.yr BP which is interpreted as a sudden change to much drier winter conditions with an increase in summer temperatures.

The point that stands out most from many of these early Holocene studies, is the dominance of forest/tree taxa during this period. This is an indication that growth conditions during the early Holocene were highly favourable for tree taxa, as in the case at Lake Mugecuo, and that climatic conditions played a major role in determining the functional diversity across the region.
4.4.2 The mid Holocene/ Holocene Optimum

The Southwest and East Asian summer monsoons have experienced changes during the Holocene, showing a sharp increase in monsoon intensity at the onset of the Holocene and reaching maxima at different times across China. According to Herzschuh (2005) regional differences in the Holocene optimum cannot be attributed to boundary shifts in the circulation systems, they are likely caused by a single climate signal (e.g. an enhanced summer monsoon in the early Holocene). In Southwestern China the maximum appeared ca. 11,000 years ago, most likely related to the maximum landward extension of the Indian summer monsoon (An et al., 2000). However, several proxies have shown (Fleitmann et al., 2003; Dykoski et al., 2005; Wang et al., 2005a; Wang et al., 2005b; Shao et al., 2006) that the EASM intensity increased sharply at the onset of the Holocene and reached a maximum intensity at ~8-9ka. The results of a study of two high-resolution palaeoclimatic records from Hidden Lake, Mozhu Gongka County (29°48.77'N, 92°32.37'E, elevation 4980 m) and Ren Co Lake, Basu County (30°43.97'N, 96°40.97'E, elevation 4450 m) in the Tibetan Plateau also revealed a maximum EAM summer monsoon in south eastern Tibet ca. 8 ka BP (Tang et al., 2000). Thereafter it reached the maximum in north western Tibet and north eastern Tibet at 7 ka BP followed by a gradually weakened monsoon system from 6 ka BP to the present (Tang et al., 2000). The first signs of the Holocene optimum appear in the Lake Mugecuo record at approximately 10.6 kcal.yr BP where *Betula* pollen reaches its highest levels of the record signifying the wettest conditions of the climatic record and thus where the EASM intensity is likely at its greatest. *Betula* abundance subsequently began to decrease at Lake Mugecuo ca. 10.4 kcal.yr BP. A decrease in mesic deciduous taxa relative to an increase in sclerophyllous taxa is an indication of the gradual decrease in spring and early summer moisture as the summer monsoon system begins to weaken through time (as seen at Lake Shayema by Jarvis, 1993) i.e. the Holocene Optimum likely ended at approximately 9.5 kal.yr BP at Lake Mugecuo. Elsewhere in the region, birches began to decline ca. 8ka BP on the eastern Qinghai-Tibetan Plateau (Ren & Beug, 2002).

A shift from *Artemisia* to Cyperaceae-dominated pollen spectra in the mid-Holocene has been inferred from pollen records on the Tibetan Plateau (Herzschuh et al., 2006a), a trend which can also be seen in the Lake Mugecuo record from ca., 5.9 kcal.yr BP. At approximately 5.9kcal.yr BP, a change in the functional diversity of the catchment vegetation appears to occur in the MG1 record with the beginnings of an increasing trend in non-arboreal pollen i.e. grasses and shrubs. A plant functional study (based on fossil pollen data) carried out at Lake Qinghai (Dallmeyer et al. 2011) on the northeastern Tibetan Plateau) also revealed a continuous decrease of forest cover during the mid-Holocene (6.0 ka) and an expansion of steppe/meadow vegetation. Whereas forests and steppe/meadow dominated the land-cover at 6.0 ka at Lake Qinghai, steppe/meadow is the prevalent biome at present-day (0.0 ka) which is dissimilar to the prevalent forest and meadow biome found at present-day Lake Mugecuo. On the whole, the seasonal and spatial patterns of temperature changes implied by the Lake Mugecuo reconstruction are consistent with previous vegetation-based estimates of the mid-Holocene climate (c.f. Shi et al., 1993; Yu et al 2000).
4.4.3 The Late Holocene

Continuous cooling and drying caused by the EAM is thought to have occurred since ca. 6000 years ago (Herzschuh et al., 2009). The gradual monsoon weakening since the mid-Holocene (as suggested by marine records), is generally interpreted as a response to declining summer insolation (Overpeck et al., 1996; Gupta et al., 2003, Herzschuh, 2006). The weakening of the Asian summer monsoon appears to have been linked more directly to decreased insolation forcing since ca. 6ka BP, which caused a cooler and drier climate in regions in the Asian monsoon region (COHMAP, 1988; Overpeck et al., 1996; Shen et al., 2005; Herzschuh, 2006; Xingqi et al., 2008). Since 3 ka BP particularly, most records from the areas dominated by the Asian monsoons show markedly lower effective moisture, while in the Westerlies-dominated areas no uniform decline in the effective moisture can be deduced (Herzschuh, 2006). A shift to increasingly more positive values in the δ\textsuperscript{18}O values in the Dongge Cave speleothem record is inferred as a decline in EAM intensity during the mid to late Holocene where it reaches its lowest intensity ca. 2.5 kcal.yr BP (Dykoski et al., 2005). Evidence of a marked shift to a cooler and drier climate can be seen in the increase in sclerophyllous taxa in the MG1 record ca. 3.6 kcal.yr BP. A climatic shift such as this would certainly be beneficial to Pinus, providing it with a competitive edge over other more temperature-sensitive taxa (Shen et al., 2005; 2006). Palaeoenvironmental and archival data from across China point to a significant climatic shift c. 3200 cal. yr BP to a cooler and/or drier climate (Zhu, 1973; Tang and Shen, 1992).

It has been shown that tree pollen content was much higher than today in pollen spectra from several areas of monsoonal central Asia during the early and mid-Holocene (e.g. Shen et al., 2005; Herzschuh, 2006). The MG1 record supports these findings as over time, increasing levels of non-arboreal taxa appear in the mid-late Holocene record. It has been found that a decline in temperatures from ~4 ka caused the decline of evergreen forests in southern China and the decline of forest cover in Tibet (Sun & Chen, 1991). An increasing trend in non-arboreal pollen from ca. 2.5 ka likely marks the contraction of forest and the expansion of meadow at Lake Mugecuo which agrees with pollen records from the eastern Tibetan Plateau and Western Sichuan Province (Sun & Chen, 1991; Ren & Beug, 2002) i.e. a notable change in the functional diversity across these sites was occurring due to the changes in prevailing climatic conditions. Deciduous oaks are not a major component of mid-elevation forests in south-western China with the exception of disturbed sites (Jarvis, 1993) which suggests that the Mugecuo catchment has been relatively undisturbed over time.

In terms of treeline changes in the catchment over time, a study by Walker (1986) in neighbouring Yunnan province of the similar (dominant) PFT types found at Lake Mugecuo indicate that these PFTs generally occur at lower altitudes in the forests of Yunnan (Figure 4.3). The vegetation types 5, 6 & 7 roughly corresponding to PFTs 5, 1 & 8 respectively in this study, are generally found at altitudes < 3000m as opposed to the approximate altitude of 3,700m at Lake Mugecuo in which representative taxa belonging to these vegetation groups currently reside. However, montane forests are common up to 4,400 m containing conifers and a broadleaved understorey (e.g. Betula) at study sites on Tibetan Plateau reviewed in Kramer et al. (2010). Therefore, it can be argued that such taxa
are found at higher altitudes in Sichuan province than in neighbouring Yunnan. In terms of the modern pollen spectra at Lake Mugecuo, warm-temperate evergreen trees are dominant and occur over relatively large areas. This type of forest is also currently found in southern and Southwestern China (from ca. 31 to 32°N latitude to the Tropic of Cancer), the northern part of Taiwan and at moderate elevations on the eastern part of the southern slope of the Tibetan Plateau (ECVC, 1980).

![Figure 4.3](image)

**Figure 4.3** Some examples of actual altitudinal zonations in the vegetation of Yunnan (after Chen, 1983). Locations: Z, Zhongdien County, NW Yunnan; W, Weixi County, NW Yunnan; J, Jindong County, west central Yunnan; M, Menghai County, south-west Yunnan; Y, Jingping County, south-central Yunnan. The local vegetation types and the general categories to which they belong are: 1, humid rainforest; 2, seasonal rainforest; 3, montane rainforest; 4, dry-valley shrubland and grassland; 5, monsoon evergreen broadleaved forest and forest of *Pinus*; 6, semi-humid evergreen broad-leaved forest and forest of *Pinus yunnanensis*; 7, montane humid evergreen broad-leaved forest; 8, mixed *Tsuga* forest; 9, montane mossy dwarf forest; 10, dry-valley small-leaved shrubland; 11, Picea and Abies forests; 12, montane shrubland and alpine meadow. Taken from *Walker* (1986).

### 4.5 Summary of overall findings

- The MG1 record provides an insight into the palaeoclimatic dynamics at Lake Mugecuo during the Late-Glacial/early Holocene transition (with evidence of the Younger Dryas) to late Holocene, indicating a long-term shift towards increasingly drier and warmer climatic conditions as the EAM weakens with some centennial fluctuations in aridity and effective moisture through time.
- This overall long-term trend is supported by the results of the PCA analysis from which three major pollen assemblages were distinguished: cool-temperate evergreen forest (*Pinus*), temperate mixed (evergreen and broadleaf) forest (*Betula*) and warm-temperate mixed (evergreen and broadleaf) forest (*Cyclobalanopsis*, *Lithocarpus*, *Taxodiaceae*).
- In terms of plant diversity, the pollen data and PFT data revealed that fluctuations in pollen diversity and changes in the plant community dynamics during the Holocene are most likely
attributed to changing climatic conditions in the catchment. The classification of taxa from the MG1 pollen spectra into plant functional types indicates a distinct shift in the dominance of certain plant communities as prevailing climatic conditions change through time. The main driver of this change was distinguished based on the PCA ordination results. The data indicated that effective moisture followed by summer temperature were the main drivers of vegetation change at Lake Mugecuo during the Holocene.

• The classification of fossil pollen data into PFTs based on a globally applicable scheme was found to have some trade-offs. The major trade-offs being that only a broad-scale view of the changes in plant community dynamics are obtained due to the static nature of the PFT classification. However, due to the nature of this study and the timescale it aimed to address, the classification utilised in this study provided an appropriate, tested, globally applicable method of assigning fossil pollen taxa to appropriate plant functional groupings which were then used in turn as a method of assessing biomization changes in the Mugecuo catchment, thus inferring broad-scale Holocene EAM variability.

• Pollen studies from this region essentially show a similar vegetation history but the boundary between these changes in often different (Jarvis, 1993; Van Campo and Gasse, 1993; Shen et al., 2005; 2006; Zhu et al., 2009) i.e. there are slight lags in the records likely due to differences in competing climatic influences across sub-regions (c.f. Cook et al., 2012). Interestingly, there are four continuous lacustrine records available for the Qinghai-Tibet plateau (Bangong Co, Fontes et al., 1996; Sumxi Co, Van Campo and Gasse,1993; Seling Co, Kashiwaya et al.,1995; and Lake Qinghai, Yu & Kelts, 2002) which show consistent trends in past changes in climate, from the Qinghai plateau to western Tibet. The most characteristic features of these records are the sudden establishment of wet conditions around 10 ka B.P., and a maximum aridity around 4-3 ka B.P which implies that the climate evolution at these four sites has been driven by the same factors. The interpretation of climatic change based upon the pollen data at Lake Mugecuo indicates a broadly similar pattern to these four sites which implies that the observed changes in palaeo-climate during the Holocene are attributable to similar palaeo-moisture regimes in this region. However, superimposed over this similar long-term trend are shorter-term climatic shifts.
Chapter 5: A reconstruction of Holocene lake catchment dynamics at Lake Mugecuo

The Lake Mugecuo record provides an insight into the environmental and climatic changes that have occurred in the catchment since c.12.6 kcal.yr BP. The two main research questions of this study were to determine a) how a dynamic high alpine lake catchment system in Sichuan Province, Southwestern China responded to major long-term changes in climatic change i.e. changes in the East Asia Monsoon since the Early Holocene and b) to reconstruct past catchment vegetation dynamics at Lake Mugecuo and past variation in sediment dynamics within the lake in order to determine the effects of plant functional diversity on catchment dynamics through time. The major temporal, physical, organic and stratigraphic features of the MG1 record are discussed in this chapter in order to address the first main research question posed in this thesis. This question is answered through the interpretation of the zoned palaeo records of %LOI, %TOC, %TN, C/N ratios and δ^13C from MG1 (presented in section 5.1, Figure 5.1). These zones provide evidence of the distinctive biogeochemical features of the MG1 record which are discussed in section 5.1. The construction of a multi-proxy zonation scheme which reveals the changes in Holocene catchment dynamics in the Mugecuo area was based on the zonation scheme used for the pollen data in order to make these proxies directly comparable in a qualitative sense. The data were zoned using this scheme as the patterns in the proxies themselves are later statistically analysed for evidence of lags in the response times of these proxies to each other. The second research aim of this study was to determine whether vegetation mediates the effects of monsoon change on past variation in lake catchment dynamics at Lake Mugecuo, a sensitive high-altitude lake. Therefore, a quantitative analysis was carried out to infer a causal link between plant functional diversity and lake catchment dynamics in the Lake Mugecuo catchment during the Holocene. This analysis is discussed in section 5.2.

The palaeo- and modern erosion rates measured from ^10Be concentrations are discussed in section 5.3. They are also discussed in relation to the organic, lithological and palaeo-vegetation proxies of the deep water core MG1 in section 5.4 as part of a comparative analysis of the multi-proxy approach to the study of catchment dynamics at Lake Mugecuo.

5.1 Discussion of organic and lithological proxies of the MG1 record

Each of the lake proxy records are discussed here as per the pollen zonation scheme outlined in section 4.2. These proxies provide evidence of organic matter and sediment that has been transported to the lake from the surrounding drainage basin as well as providing the evidence to distinguish it from organic matter that is produced/formed within the lake itself. The organic and lithological proxies are outlined in Figure 5.1.
Figure 5.1 Zoned organic and lithological proxies of MG1 record with modern and palaeo-erosion rates (discussed in later sections), sedimentation rates between dated points and the smoothed D4 δ¹⁸O speleothem data are included for reference.
5.1.1 Zone MPZ -1 (12.6- 11.9 kcal.yr BP; 383-322cm)

This basal multi-proxy zone spans c. 700 years. Silt-clay rich gyttja dominates this section of the MG1 record. The low levels of sand in the core coupled with a dominance of silt-clay rich gyttja suggests that low velocity flows (c.f. Knighton, 1984) were in operation within the lake and lake catchment during this period. The %LOI, %TOC and %TN data reveal a general increasing trend in organic matter and organic productivity in this zone. The %LOI values in this zone range from <10% up to > 20% indicating increasing organic sediment input to the basin (c.f. Birks and Birks, 2006). The coarse resolution of this proxy makes it difficult to determine whether these fluctuations are indicative of short-term changes i.e. decadal or longer-term changes i.e. centennial. Although there are tantalising hints of centennial-scale changes in %LOI, the age control remains insufficient to make the identification of such events with confidence. The analysis of results was therefore restricted to the interpretation of broad trends in the data. The %TOC values indicate fluctuations in organic productivity with values ranging from 1.5-9.5%. The fluctuations in %TOC may represent changes in the origin of organic matter and/or changes in depositional processes during this period (c.f. Meyers & Lallier-Vergès, 1999). The %TN values are generally low, <0.8% indicating there is little organic nitrogen present, but an increasing trend may indicate increasing organic activity originating from plants and soil sources. Increasing %TN values are sustained during this period before dropping dramatically to lows of 0.3% at approximately 12.0 kcal.yr BP. C/N values range from 7.5 to 10.5 suggesting that aquatic plants are the primary contribution of OM content to the lake archive (Meyers & Lallier-Vergès, 1999). δ¹³C values vary considerably throughout this zone from c. -24 to -20‰. Less negative δ¹³C values may be representative of a phase of organic productivity driven by a mixture of terrestrial and aquatic processes i.e. a change in pedogenesis caused by an alternative source of carbon and nitrogen (other than the atmosphere) (Herzschuh et al., 2005). However, δ¹³C values drop dramatically at approximately 12.2 kcal.yr BP indicating a rapid shift to increased aquatic organic productivity for the remainder of this zone.

On the whole, it appears that increasing levels of organic matter were archived in the lake in the basal zone of the MG1 record but the majority of this organic matter was aquatic in origin. Estimated sedimentation rates during this period stand at approximately 0.1cm/yr. Although sedimentation rates were calculated based on the depths between dated points and thus are a rough proxy of sediment influx to the lake basin, there is little evidence of any notable erosion events in the catchment and subsequent deposition into the lake based on the other lake proxies in this section of the core. Thus sedimentation rates are interpreted here as a reasonable proxy for landscape change in this instance.

X-ray diffraction (XRD) mineralogical analysis of the organic clay-rich gyttja bands found in the MG1 core revealed high levels of illite (70-90%). High levels of illite in clay minerals are interpreted as corresponding to relatively dry periods as per Singer (1984). A mineral assemblage containing high levels of illite in the clay size fraction has frequently been observed in soils and sediments produced by high-latitude or cold-climate weathering (Bockheim, 1982; Campbell and Claridge, 1982). Thus,
illite is inferred here to be indicative of weak weathering intensities, a cold and dry environment in the source area, with physical weathering dominant (Singer, 1984). These climatic conditions are similar to those inferred from the MG1 palaeoecological record in section 4.1.2 during this period which are also found at other sites across the region (Jarvis, 1993; Herzschuh, 2006).

5.1.2 Zone MPZ 2 (11.9-8.3 kcal.yr BP; 322-232cm)

Spanning approximately 3,600 years, this zone is characterised by a decline in sedimentation rates at approximately 9.9 kcal.yr BP and a general increasing dominance of silt particles in the sediment size record. The increased deposition of fine silt particles suggests that lake energy remained low in this section of the core. However, sand levels reach their highest of the core of 3% at 9.4 kcal.yr BP (269cm) which coincides with a peak in total pollen concentrations (Figure 4.1) suggesting an influx of sediment to the lake. Following this, mixed laminations of darker and lighter clastic sediment from approximately 9.2 kcal.yr BP indicate a period of enhanced bottom-current activity and/or enhanced sediment inwash to the lake (as per Lister et al., 1991). The δ¹³C values shift dramatically in the early section of this zone reaching highs of approximately -20. Increases in %TOC and %TN and less negative δ¹³C values indicate a phase of organic productivity driven by terrestrial and aquatic processes. Higher C/N values than the previous zone suggest an increase of terrestrially derived organic carbon entering the lake. However, a notable change in the OM proxies is seen from approximately 9.1-9.0 kcal.yr BP where C/N levels drop and increasingly negative δ¹³C values indicate a shift to a definite dominance of aquatic organic productivity in the lake.

5.1.3 Zone MPZ 3 (8.3-5.8 kcal.yr BP; 232-172cm)

This zone spans approximately 2,500 years. A low C/N ratio (less than 10) in the early portion of this zone (8.3-7.0 kcal.yr BP) suggests that the organic component of the MG1 sediment record is primarily aquatic in origin. Mixed laminations of darker and lighter clastic sediment are evident in the MG1 stratigraphic record until approximately 7.0 kcal.yr BP. Sand levels in the core drop dramatically at this point which suggests that the mixed laminations are due to enhanced sediment inwash which is evidenced by the coarse sediment component in the particle size record. Sedimentation rates increase dramatically at approximately 7.1 kcal.yr BP which coincides with elevated %LOI values likely indicating a slight influx of organic material into the lake. Elevated C/N values point to a sub-equal mixture of aquatic and terrestrial organic productivity during this period. The dominance of silt particles and general increasing trend seen in the previous zone continues in this section of the core until approximately 6.2 kcal.yr BP. Silt values plummet at this point to their lowest values recorded in the MG1 record (58-60%) with clay particle size values reaching their highest of the record of almost 40%. A band of organic clay-rich gyttja indicates a cold and dry physical weathering environment at this point (see explanation of high levels of illite above). This corresponds to the climatic conditions inferred from the pollen data for this period which point to a period of cold, dry winter conditions prevailing. Declining organic productivity within the lake catchment is apparent based on declining %TOC and %TN values. The organic matter (OM) proxy records suggest that the majority of organic
matter is derived from lacustrine sources from 9.2-5.8 kcal.yr BP. Although a trend of slightly less negative $\delta^{13}C$ values occurs throughout this zone, the $\delta^{13}C$ data generally points to a lacustrine environment that is predominantly influenced by aquatic plant organic matter.

5.1.4 Zone MPZ 4 (5.8-2.3 kcal.yr BP; 172-85 cm)

Spanning approximately 3,500 cal. years, this zone is characterised by the fastest levels of sedimentation rates of the entire MG1 record. Based on mean (smoothed) sedimentation rates between dated samples, sedimentation rates range from approximately 0.06-0.2 cm/yr in this section of the MG1 record. The increased rates of up to 0.2 cm/yr at approximately 2.4 kcal.yr BP suggest an increase in availability and/or sediment supply during which may be related to a decline in arboreal pollen during this period (see Figure 4.1).

The core stratigraphy indicates a shift from organic clay-rich gyttja to mixed fine laminations of darker and lighter clastic sediment from approximately 5.8-4.1 kcal.yr BP. A period of increased sand particle size values from approximately 5.0-4.1 kcal.yr BP suggests that this was a period of increased erosion in the lake catchment which resulted in increased sediment in-wash into the lake. From 5.8-4.2 kcal.yr BP an increase in sedimentation rates coincides with an increase in OM productivity which may signify the influx of material into the lake from the surrounding catchment due to increased erosion in the catchment. The MG1 stratigraphic and particle size records show that this was followed by a band of organic clay-rich gyttja (4.1-3.7 kcal.yr BP) indicating a transition to cold and dry physical weathering environmental conditions with low velocity flows into the lake as evidenced by low values of coarse particle size sediments. From approximately 4.0 kcal.yr BP, the MG1 organic matter proxies also show a decline in productivity within the lake with a decline in %TOC and %TN. In terms of the overall trend of the OM proxies for this zone, C/N values range from approximately 8.0-10.5 and $\delta^{13}C$ values remain low thus indicating that the majority of organic matter in the lake is aquatic in origin.

5.1.5 Zone MPZ 5 (2.3-0 kcal.yr BP; 85-0 cm)

This upper-most section of the MG1 record spans approximately 2,300 years. For the first time in the record a sustained period of increased sand levels in the particle size record occurs from approximately 2.2-1.0 kcal.yr BP (78-30 cm). Levels of sand reach highs of almost 3% during this period. A shift from primarily silt/clay sediments (associated with a lower-energy environment and/or deeper waters) to higher levels of sand (associated with a higher-energy environment and/or shallower waters) could indicate that the shoreline was closer (e.g. Mischke et al., 2005) to the MG1 coring site (see Figure 2.11) implying a shift to lower lake stands during this period. A change such as this would likely trigger a change in the abundances of aquatic plants growing at the lake edges during this period (Wilcox and Meeker, 1991). However, the MG1 pollen record indicates that there were no significant shifts in aquatic pollen at this time. It is therefore more likely that this shift to increased levels of sand reflect changes in lake energy and/or sediment transport dynamics.
Fluctuations in %TOC and %TN indicate fluctuations in organic matter productivity throughout this upper section of the MG1 record. On the whole, δ^{13}C values remain low and C/N values remain below 10.5 aside from a peak of approximately 12 at 2.2 kcal.yr BP. A notable shift in C/N ratios at approximately 1.2kcal.yr BP likely represents an increased dominance of lacustrine vegetation contributing to organic productivity within the lake. This coincides with a decline in sedimentation rates which suggests a lower influx of sediment into the lake which is bringing little terrestrial organic matter with it. According to Mischke et al. (2005), a high terrigenous proportion of organic matter in lake sediments may arise from an overall increased loading of air or river water by sediments and/or simply may reflect shorelines nearer to the core site and thus a lake with a reduced depth and area. The generally low terrigenous proportion of organic matter in the entire MG1 record and the lack of coarse sand material throughout may be indicative of a relatively stable lake level through time.

5.2 A statistical comparison of the MG1 palaeo-vegetation record and the MG1 lake catchment dynamics proxy records

One of the primary aims of this study was to reconstruct past catchment vegetation dynamics at Lake Mugucuo and past variation in sediment dynamics within the lake in order to determine the effects of plant functional diversity on catchment dynamics through time. A preliminary analysis of the different proxies traditionally utilised in palaeoenvironmental reconstructions (e.g. Leeder et al., 1992; Jarvis, 1993a; Sperazza et al., 2004; Birks & Birks, 2005; Leng et al., 2006) adopted in this study have been thus far discussed with a view to compare the changes in catchment dynamics and plant community dynamics inferred from these proxies through time. In order to infer causality in the relationship between plant community dynamics and the hypothesized concurrent changes in lake catchment dynamics, further quantitative analysis was necessary to determine whether a statistically significant relationship exists between them based on the data obtained there. To address this issue, analysis of the proxy data focused on the plant functional types (absolute count data) which were used as the primary indicator for changes in functional diversity through time whilst the particle size data was used as the primary indication of changes in landscape cover i.e. catchment dynamics through time. Unfortunately it was not possible to obtain a whole Holocene record of ^{10}Be erosion rate measurements due to time constraints of lake sediment sampling at the catchment site as previously mentioned in section 1.6.

Pollen assemblages (Leopold et al., 1982 and Kutzbach et al., 1996) and grain-size measurements (Halfman and Thomas, 1984 and Noren et al., 2002) are widely used in global lake sediment studies. Within a lake, shallow water areas close to the shoreline are higher energy zones relative to deeper areas which are further away from the shoreline. This results in the deposition of larger (sand) particles close to the shore line, and the deposition of smaller (silt and clay) particles in the deeper parts of the lake (Håkanson & Jansson, 1983; Leeder et al., 1992). Therefore, particle size data can reveal a lot about the delivery of sediment into a lake whilst providing an insight into catchment stability and erosion rates (Loveland & Whalley, 2000; Sperazza et al., 2004). The (absolute)
measurements of five clastic particle sizes (PS < 4.00 μm, PS >4μm < 16μm, PS >16μm <32 μm, PS >32μm <64μm, PS >64 μm) within the lake MG1 sediment record were therefore used as a proxy for fluvial dynamics and sediment deposition/transport i.e. lake catchment dynamics. The D4 δ18O record from Dongge Cave (presented in section 2.8) was used in this analysis as a guide to the pattern of changes in East Asian Monsoon (EAM) intensity for the Holocene. This proxy was utilised in order to test the main research hypothesis:

‘Does vegetation mediate the effects of climatic change on lake catchment dynamics during the Holocene at Lake Mugecuo?’

The inter-comparison and critical assessment of the three proxy records in relation to each other was achieved by normalising all curves and placing them on the same axis. These data were de-trended, centred and expressed in terms of standard deviation units. The quantitative analysis of the data focussed on the following hypotheses:

1. Is there a significant correlation between climate and plant functional diversity in the lake Mugecuo catchment?
2. Is there a significant correlation between changes in plant functional diversity and changes in catchment dynamics in the lake Mugecuo catchment?
3. Is there a significant correlation between climate and catchment dynamics in the Lake Mugecuo record?
4. Is there a causal relationship between changes in plant functional diversity and changes in lake catchment dynamics in the lake Mugecuo catchment (regression analysis)?
5. Plant functional diversity does not mediate the effects of climate on lake catchment dynamics (regression analysis)

In accepting hypotheses 3 and 4, it could reasonably be inferred that plant functional diversity is a driver of lake catchment dynamics and mediates the effects of climate on catchment dynamics. Each of these hypotheses was tested using an array of robust statistical analysis techniques. Due to the complex nature of this analysis and the large volume of output data, the main conclusions drawn from this testing are outlined in the following paragraph for the benefit of the reader. A brief synopsis of each of the techniques is then presented along with the supporting data that led to each of the main conclusions. A comparative discussion and critical analysis of the results is then discussed in section 5.2.6. Supporting output data are provided in Appendix A.2.

### 5.2.1 Summary of Statistical Analysis Findings

- Plant functional diversity is more correlated to lake catchment dynamics on shorter timescales than climate is to lake catchment dynamics i.e. plant functional diversity interact with landscape dynamics on a shorter-term timescale in the Mugecuo catchment <2,800 years.
• A causal relationship between plant functional diversity and the four of the five particle size (lake catchment dynamics proxies) in the MG1 record was shown.
• It is inferred from these results that vegetation likely mediates the effects of climate change on lake catchment dynamics at Lake Mugecuo.
• Causal relationships were found between plant functional diversity and lake catchment dynamics when several PFTs were included in the model.
• Analysis indicates that shrubs are the most important variables in relation to changes in coarse particle size in the catchment i.e. they more highly correlated to larger erosion/denudation events.
• Changes in the finer sediment sizes appear to be influenced by an array of plant functional diversity types e.g. sedges and grasses

5.2.2 Spearman Rank Correlation

A simple Spearman Rank correlation coefficient was calculated as a preliminary test for correlations between the climate, plant functional type and particle size data. Significant correlations for each category of particle size data are outlined in Table 5.1. The results of the Spearman correlation coefficient revealed that significant correlations were found between PFT3 (sedges) and particle size (PS) category < 4.00 μm, PFT1 (Pinus and Tsuga) and PS category >4μm < 16μm, PFT3 and PS category >16μm <32 μm, PFT1 & PFT 7 (Artemisia) and PS category >32μm <64μm and PFT1 & PFT3 and PS category >64μm. The most significant correlations found between climate and plant functional diversity (Table 5.2) were PFT1 (Pinus and Tsuga), PFT 5 (Betula and Rosaceae) and PFT 3 (Cyperaceae), whilst climate was most significantly (negatively) correlated to PS >16μm <32 μm. These results indicated that further analysis was required to delve deeper into the relationships between these proxies in order to test the hypotheses outlined previously. Note: Supplementary data of the Spearman correlation coefficients of the Dongge Cave speleothem data versus the PS data are presented in Appendix A.2.1.
Table 5.1 Spearmans correlation coefficients ($R^2$) for each of the nine plant functional types (count data) and the five particle size data categories. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.

<table>
<thead>
<tr>
<th>PFT</th>
<th>PS &lt; 4.00 µm</th>
<th>PS &gt;4µm</th>
<th>PS &gt;16µm</th>
<th>PS &gt;32µm</th>
<th>PS &gt;64µm</th>
</tr>
</thead>
<tbody>
<tr>
<td>PFT1 (Pinus, Tsuga)</td>
<td>-0.2627</td>
<td>0.4217</td>
<td>-0.0375</td>
<td>-0.2566</td>
<td>-0.3918</td>
</tr>
<tr>
<td>PFT2 (Poaceae)</td>
<td>0.2802</td>
<td>-0.4197</td>
<td>-0.0081</td>
<td>0.2271</td>
<td>0.2104</td>
</tr>
<tr>
<td>PFT3 (Cyperaceae)</td>
<td>0.5133</td>
<td>-0.0195</td>
<td>-0.5534</td>
<td>-0.0439</td>
<td>0.3838</td>
</tr>
<tr>
<td>PFT4 (Carpinus)</td>
<td>0.2170</td>
<td>0.0752</td>
<td>-0.2314</td>
<td>-0.2186</td>
<td>0.0979</td>
</tr>
<tr>
<td>PFT5 (Betula, Rosaceae)</td>
<td>0.1775</td>
<td>-0.3620</td>
<td>0.0613</td>
<td>0.2400</td>
<td>0.2165</td>
</tr>
<tr>
<td>PFT6 (Thalictrum)</td>
<td>0.1344</td>
<td>0.0267</td>
<td>-0.0904</td>
<td>0.0053</td>
<td>0.1290</td>
</tr>
<tr>
<td>PFT7 (Artemisia)</td>
<td>0.2397</td>
<td>-0.2420</td>
<td>-0.0342</td>
<td>0.2453</td>
<td>0.3223</td>
</tr>
<tr>
<td>PFT8 (Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae)</td>
<td>-0.0966</td>
<td>0.0141</td>
<td>0.1710</td>
<td>-0.0222</td>
<td>0.1124</td>
</tr>
<tr>
<td>PFT 9 (Ericaceae)</td>
<td>0.2186</td>
<td>0.1875</td>
<td>-0.4363</td>
<td>-0.0263</td>
<td>0.0935</td>
</tr>
</tbody>
</table>

Table 5.2 Spearmans correlation coefficients ($R^2$) for each of the nine plant functional types (count data) and the Dongge Cave speleothem data. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.

<table>
<thead>
<tr>
<th>Proxy</th>
<th>D4 $\delta^{18}O$</th>
</tr>
</thead>
<tbody>
<tr>
<td>PFT1 (Pinus, Tsuga)</td>
<td>0.42</td>
</tr>
<tr>
<td>PFT2 (Poaceae)</td>
<td>-0.04</td>
</tr>
<tr>
<td>PFT3 (Cyperaceae)</td>
<td>0.31</td>
</tr>
<tr>
<td>PFT4 (Carpinus)</td>
<td>0.19</td>
</tr>
<tr>
<td>PFT5 (Betula, Rosaceae)</td>
<td>-0.40</td>
</tr>
<tr>
<td>PFT6 (Thalictrum)</td>
<td>0.02</td>
</tr>
<tr>
<td>PFT7 (Artemisia)</td>
<td>-0.14</td>
</tr>
<tr>
<td>PFT8 (Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae)</td>
<td>-0.14</td>
</tr>
<tr>
<td>PFT 9 (Ericaceae)</td>
<td>0.23</td>
</tr>
</tbody>
</table>
5.2.3 Cross-Correlation Analysis

Cross-correlation analysis quantifies the similarity between two waveforms (signals) as a function of a time-lag applied to one of the signals. Cross-correlation was used to test whether it takes some unknown number of years for each of the three main proxies to respond to each other i.e. climate, plant functional diversity and catchment dynamics. Cross correlation was performed using the xcorr function in Matlab v. R2011 which returns a cross-correlation sequence over a lag range. Significance levels were calculated using a Monte Carlo approach which involves randomising the data, calculating the cross-correlation and iterating it 5,000 times in order to 'respect' the structure of the data whilst exploring the full range of permutations that might arise by random chance (Jones et al., 2013).

Cross-correlation analysis was carried out to quantify the similarity between the Dongge Cave speleothem data signal and the particle size data signals in order to test the hypothesis: there is no significant lag in the response of lake catchment dynamics in the Lake Mugecuo record to changes in the EAM during the Holocene. The results indicate (Appendix A.2.2) that weak but significant lagged landscape responses are observed on millennial timescales between the D4 δ18O data and the particle sizes < 4.00 μm, >4μm < 16μm, >16μm <32 μm, >32μm <64μm and >64μm. This implies that the particle size data (i.e. catchment dynamics) respond to long-term (103) changes in climate on millennial timescales in the Lake Mugecuo catchment during the Holocene. A Lomb-Scargle Fourier transform (Lomb, 1976; Scargle, 1982; Savaransky, 2008) (which computes the Lomb normalized periodogram/spectral power as a function of frequency) of the climate data revealed that long-term periodicities do in fact dominate the speleothem data (Figure 5.2, Appendix A.2.2). The Lomb-Scargle Fourier transform indicates that the long-term (103) trends identified by cross-correlation in the smoothed D4 δ18O record are real trends in the data and not just spurious peaks. In a Lomb-Scargle Fourier transform, the significance returned is the probability of a false positive type 1 error. Low probability values which can be seen in Figure 5.2 when there is a peak in frequency indicate a high degree of significance in the associated periodic signal.

Figure 5.2 Lomb-Scargle Fourier periodogram of the regional climate record (in blue) inferred from the D4 δ18O at Dongge Cave. The estimated significance of the power values are presented on a secondary axis (probability in green).
Cross-correlation analysis was carried out to quantify the similarity between the Dongge Cave speleothem data signal and the plant functional type data signals in order to test whether there is a significant lag in the response of plant functional diversity in the Lake Mugecuo record to changes in the EAM during the Holocene. The results indicated (Appendix A.2.2) that weak but significant lagged landscape responses are observed on millennial timescales between the D4 δ¹⁸O data and the nine plant functional types. This implies that plant functional diversity responds to long-term changes in climate on millennial timescales in the Lake Mugecuo catchment during the Holocene.

Cross-correlation analysis was carried out to quantify the similarity between the plant functional type data signals and each of the particle size fraction signals in order to test whether there is a significant lag in the response of lake catchment dynamics in the Lake Mugecuo record to changes in the plant functional diversity during the Holocene. On the whole, the results indicate (Appendix A.2.2) that weak but significant lagged landscape responses are observed on millennial timescales between the particle size fractions <4µm, >4µm < 16µm, and >16µm <32 µm and certain plant functional type groupings. This implies that these particle size data (i.e. catchment dynamics) respond to long-term changes in plant functional diversity on millennial timescales in the Lake Mugecuo catchment during the Holocene. The overriding trends in these particle size data were analysed using Lomb-Scargle Fourier transform which revealed that the cross-correlations of these data with the plant functional time-series data did yield valid results. The Lomb-Scargle periodograms for these data show that long-term trends (10³) dominate these time-series (Figure 5.3), with high degrees of significance in the associated periodic signals.

Figure 5.3 Lomb-Scargle Fourier periodogram of the particle size categories <4µm, >4µm < 16µm, and >16µm <32 µm (in blue). The estimated significance of the power values are presented on a secondary axis (probability in green).
The cross-correlation results indicated (Figure 5.4) that particle sizes >32µm <64µm and >64µm display lags on centennial scales in response to certain PFTs. Particle size fraction >32µm <64µm displays a centennial-scale lag of 264 years when cross-correlated with PFT2 (Poaceae), while particle size fraction >64µm displays a centennial-scale lag of 198 years when cross-correlated with PFT2 (Poaceae) and a centennial-scale lag of 264 years when cross-correlated with PFT7 (Artemisia).

Figure 5.4 Cross-correlation showing the lagged response of PS >32µm <64µm & PS >64µm on centennial scales to certain PFTs. PS >32µm <64µm displays a centennial-scale lag of 264 years when cross-correlated with PFT2 (Poaceae). PS >64µm displays a centennial-scale lag of 198 years when cross-correlated with PFT2 (Poaceae) and a centennial-scale lag of 264 years when cross-correlated with PFT7 (Artemisia). Significance (in red) is denoted by 95% confidence lines beyond which correlations are unlikely to arise purely by chance.

In order to determine which, if any of the catchment dynamics proxies (i.e. particle size classes) were significantly correlated to the long-term plant functional diversity record (i.e. the nine plant functional type groupings), a cross-spectrum analysis was performed between each of the particle size proxies and the PFT groupings.
5.2.4 Cross-spectral analysis

Cross-spectral analysis enables one to ‘correlate’ between frequency peaks in the auto-spectrum of each time-series (Jones et al., 2013). Cross-spectral analysis was used to test whether the frequency peaks of the three proxy datasets i.e. climate, plant functional diversity and catchment dynamics significantly correlate with each other. The Matlab v. R2011 function cpsd estimates the cross power spectral density $P_{xy}$ of the discrete-time signals $x$ and $y$ using the Welch's averaged, modified periodogram method of spectral estimation. The cross power spectral density is the distribution of power per unit frequency. Where centennial-scale lagged responses were found in the cross-correlation analysis e.g. particle size fractions >32µm <64µm and >64µm (see section 5.2.3), the data was adjusted for these lags (i.e. the offset was removed) in order to make the spectral series directly comparable. Significance levels were calculated using a Monte Carlo approach which involves randomising the data, calculating the cross-spectra and iterating it 5,000 times in order to ‘respect’ the structure of the data whilst exploring the full range of permutations that might arise by random chance (Jones et al., 2013).

Cross-spectral analysis was carried out to quantify the similarity between the Dongge Cave speleothem data signal and the plant functional type data signals (Figure 5.5) in order to test whether there is a significant correlation between climate and plant functional diversity in the lake Mugecuo catchment. The results indicate that the dominant correlating periodicities are found between climate and PFT 2, PFT 4, PFT 6 (2,850 years) and PFT 9 (3,379 years).

Figure 5.5 Cross-spectra between plant functional diversity and the D4 $\delta^{18}$O record at Dongge Cave. All data was detrended, centred and expressed in terms of standard deviation units. Only those cross-spectra that proved significant are shown here. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-spectral analysis was carried out to quantify the similarity between the Dongge Cave speleothem data signal and the particle size data signals (Figure 5.6) in order to test whether there is a significant correlation between climate and catchment dynamics in the Lake Mugecuo record.

The results indicate that the dominant correlating periodicities are found between climate and PS >4µm < 16µm (~2,800 years), >32µm <64µm (5,600 years) and >64µm (~2,800 years).

![Cross-spectra between the particle size fractions >4µm < 16µm, >32µm <64µm and >64µm and the Dongge Cave D4 δ¹⁸O record](image)

The dominant periodicities are found between PS >4µm < 16µm, >32µm <64µm and >64µm and PFT 4 (Carpinus; 1,445 years), PFT 2 (Poaceae; 1,580 years), PFT 3 (Cyperaceae; 2,111 years) and PFT 9 (Ericaceae; 2,413 years). In the case of particle size >16µm <32 µm, there were no particularly dominant correlating periodicities with the plant functional type groupings (see Appendix A.2.4). For particle size >32µm <64µm, the dominant correlating periodicities are found between PFT1
(Pinus & Tsuga; 1,207 years), PFTs 4 & 6 (Thalictrum; 1,408 years) and PFT 2 (2,413 years). For particle size >64µm, the dominant correlating periodicities are found between PFT 8 (1,877 years).

Figure 5.7 Cross-spectra between the particle size fraction < 4.00 µm and the plant functional types 4 & 8. All data was detrended, centred and expressed in terms of standard deviation units. Only those cross-spectra that proved significant are shown here. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.

Figure 5.8 Cross-spectra between the particle size fraction >4µm < 16µm and plant functional diversity. All data was detrended, centred and expressed in terms of standard deviation units. Only those cross-spectra that proved significant are shown here. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
Figure 5.9 Cross-spectra between the particle size fraction >32µm <64µm and plant functional diversity. All data was detrended, centred and expressed in terms of standard deviation units. Only those cross-spectra that proved significant are shown here. For PFT 2 a lag of 264 years was removed from the data. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.

Figure 5.10 Cross-spectra between the particle size fraction >64µm and plant functional diversity. All data was detrended, centred and expressed in terms of standard deviation units. Only those cross-spectra that proved significant are shown here. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
5.2.4.1 A brief summary of the statistical analysis results so far

- Plant functional diversity responds to long-term changes ($10^3$) in climate on millennial timescales in the Lake Mugecuo catchment during the Holocene.
- The particle size data respond to long-term ($10^3$) changes in climate on millennial timescales in the Lake Mugecuo catchment during the Holocene.
- The particle size fractions <4µm, >4µm < 16µm, and >16µm <32 µm respond to long-term changes in plant functional diversity on millennial timescales.
- PS >32µm <64µm was found to display a centennial-scale lag of 264 years when cross-correlated with PFT2 (Poaceae)
- PS >64µm displays a centennial-scale lag of 198 years when cross-correlated with PFT2 (Poaceae) and a centennial-scale lag of 264 years when cross-correlated with PFT7 (Artemisia).
- Correlating periodicities were found between PS <4µm and PFTs 4 & 8.
- Correlating periodicities were found between PS >4µm < 16µm and PFTs 2, 3, 4, 9 and the climate data.
- No dominant correlating periodicities were found between PS >16µm <32 when tested against the climate and PFT data.
- Correlating periodicities were found between PS >32µm <64µm and PFTs 1, 2, 4 & 6 and the climate data.
- Correlating periodicities were found between PS <64µm and PFT8 and the climate data.

*Note: All of the correlating periodicities were on millennial timescales.

5.2.5 Stepwise regression analysis

The next logical step of the analysis was to determine a causal relationship between the particle size data and the plant functional type groupings. This analysis was therefore followed by a stepwise multiple regression with forward selection of variables in order to test whether there is a causal relationship between changes in plant functional diversity and changes in lake catchment dynamics in the lake Mugecuo catchment. Stepwise regression is a systematic method for adding and removing terms from a multi-linear model based on their statistical significance in a regression. Stepwise multiple regression with forward selection of variables was used in this analysis which involves starting with no variables in the model. The model then works by testing the addition of each independent variable using a chosen model comparison method, adding the independent variable (if any) that improves the model the most, and repeating this process until none improves the model. At each step, the p value of an F-statistic is computed to test models with and without a potential term. If a term is not currently in the model, the null hypothesis is that the term would have a zero coefficient if added to the model. If there is sufficient evidence to reject the null hypothesis, the term is added to the model. The dependent variables i.e. the response variables in this case were the five particle size
fractions. The independent variables tested in the multiple regression analysis were the plant functional diversity proxies i.e. PFTs 1-9. Variables were selected in the multiple regression only if P, 0.001. The stepwise regression analysis was then repeated testing the five particle size categories against the individual pollen taxa/species data which were used to create the plant functional type groupings in order to test whether any single plant taxa was the primary driving force behind changes in landscape cover at the Lake Mugecuo catchment. Significant results are included/discussed where appropriate. Stepwise regression was carried out using the stepwise function in Matlab v. R2011 (see Appendix A.2.5 for script details).

5.2.5.1 Stepwise Regression Particle Size <4 μm

Stepwise regression analysis was used to determine if plant functional diversity was a driver of changes in the particle size class < 4.00 μm (dependent variable) in the Lake Mugecuo during the Holocene. The model showed (Table 5.3) that PFT3 (Cyperaceae) was the most correlated variable with PS class < 4.00 μm ($R^2$=0.285) in conjunction with PFT2, PFT5 and PFT 9 (Figure 5.11) i.e. there is a causal relationship between PFT 3 and PS < 4.00 μm but only when these other PFTs are taken into account in the model.

Table 5.3 Multiple regression analysis of particle size class < 4.00 μm (dependent variable) as a function of functional diversity (plant functional types). The co-dependent variables are highlighted in bold.

<table>
<thead>
<tr>
<th>Regression Variables</th>
<th>Partial Regression Coefficient</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>PFT1 (Pinus, Tsuga)</td>
<td>-0.046</td>
<td>0.5642</td>
</tr>
<tr>
<td>PFT2 (Poaceae)</td>
<td>0.144</td>
<td>0.0313</td>
</tr>
<tr>
<td>PFT3 (Cyperaceae)</td>
<td><strong>0.285</strong></td>
<td>0.0000</td>
</tr>
<tr>
<td>PFT4 (Carpinus)</td>
<td>0.115</td>
<td>0.0955</td>
</tr>
<tr>
<td>PFT5 (Betula, Rosaceae)</td>
<td>0.145</td>
<td>0.0326</td>
</tr>
<tr>
<td>PFT6 (Thalictrum)</td>
<td>0.023</td>
<td>0.7202</td>
</tr>
<tr>
<td>PFT7 (Artemisia)</td>
<td>0.014</td>
<td>0.8355</td>
</tr>
<tr>
<td>PFT8 (Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae)</td>
<td>-0.078</td>
<td>0.2342</td>
</tr>
<tr>
<td>PFT 9 (Ericaceae)</td>
<td>0.200</td>
<td>0.0024</td>
</tr>
</tbody>
</table>

Notes: Model Summary $R^2$=0.207, adjusted $R^2$=0.190; P < 0.0001. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.
Partial regression leverage plot for particle size class < 4.00 μm (dependent variable) as a function of PFT3 (Cyperaceae). The best model for the data was achieved by adding the independent variables PFT2, PFT5 and PFT 9.

5.2.5.2 Stepwise Regression Particle Size >4µm < 16µm

The results of the stepwise model for particle size class >4µm < 16µm (dependent variable) as a function of functional diversity (plant functional types) revealed that PFT1 (Pinus and Tsuga) was most correlated to this particle size fraction with adjustments for PFT 2 (Poaceae), PFT 4 (Carpinus) and PFT 9 (Ericaceae) (Figure 5.12, Table 5.4). A weak but significant negative correlation was found between PFT 2 and particle size class >4µm < 16µm.
Table 5.4 Multiple regression analysis of particle size class >4µm < 16µm (dependent variable) as a function of functional diversity (plant functional types). The co-dependent variables are highlighted in bold.

<table>
<thead>
<tr>
<th>Regression Variables</th>
<th>Partial Regression Coefficient</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>PFT1 (Pinus, Tsuga)</td>
<td>0.249</td>
<td>0.0000</td>
</tr>
<tr>
<td>PFT2 (Poaceae)</td>
<td>-0.339</td>
<td>0.0000</td>
</tr>
<tr>
<td>PFT3 (Cyperaceae)</td>
<td>0.142</td>
<td>0.0841</td>
</tr>
<tr>
<td>PFT4 (Carpinus)</td>
<td>0.124</td>
<td>0.0179</td>
</tr>
<tr>
<td>PFT5 (Betula, Rosaceae)</td>
<td>-0.153</td>
<td>0.2525</td>
</tr>
<tr>
<td>PFT6 (Thalictrum)</td>
<td>0.070</td>
<td>0.1745</td>
</tr>
<tr>
<td>PFT7 (Artemisia)</td>
<td>0.072</td>
<td>0.2531</td>
</tr>
<tr>
<td>PFT8 (Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae)</td>
<td>-0.048</td>
<td>0.5709</td>
</tr>
<tr>
<td>PFT 9 (Ericaceae)</td>
<td>0.112</td>
<td>0.0340</td>
</tr>
</tbody>
</table>

Notes: Model Summary $R^2=0.344$, adjusted $R^2=0.329$; $P < 0.0001$. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.

5.2.5.3 Stepwise Regression Particle Size >16µm <32 µm

Stepwise regression analysis was used to determine if plant functional diversity has driven changes in the particle size class >16µm <32 µm (dependent variable) in the Lake Mugecuo during the Holocene. The model showed (Table 5.4) that PFT3 (Cyperaceae) was the most correlated variable with PS class >16µm <32 µm when adjusted for PFT4, PFT8 and PFT 9 (Figure 5.12) i.e. there is a causal relationship between PFT 3 and PS >16µm <32 µm but only when these other PFTs are taken into account in the model.
Figure 5.13 Partial regression leverage plot for particle size class >16µm <32 µm (dependent variable) as a function of Cyperaceae (PFT3). The best model for the data was achieved by adding the independent variables PFT4, PFT8 and PFT 9.

Table 2 Table 5.5 Multiple regression analysis of particle size class >16µm <32 µm (dependent variable) as a function of functional diversity (plant functional types). The co-dependent variables are highlighted in bold.

<table>
<thead>
<tr>
<th>Regression Variables</th>
<th>Partial Regression Coefficient</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>PFT1 (Pinus, Tsuga)</td>
<td>0.096</td>
<td>0.1648</td>
</tr>
<tr>
<td>PFT2 (Poaceae)</td>
<td>0.023</td>
<td>0.7595</td>
</tr>
<tr>
<td>PFT3 (Cyperaceae)</td>
<td><strong>-0.567</strong></td>
<td>0.0000</td>
</tr>
<tr>
<td>PFT4 (Carpinus)</td>
<td>-0.269</td>
<td>0.0009</td>
</tr>
<tr>
<td>PFT5 (Betula, Rosaceae)</td>
<td>-0.110</td>
<td>0.4167</td>
</tr>
<tr>
<td>PFT6 (Thalictrum)</td>
<td>-0.114</td>
<td>0.0876</td>
</tr>
<tr>
<td>PFT7 (Artemisia)</td>
<td>-0.118</td>
<td>0.0705</td>
</tr>
<tr>
<td>PFT8 (Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae)</td>
<td><strong>0.420</strong></td>
<td>0.0000</td>
</tr>
<tr>
<td>PFT 9 (Ericaceae)</td>
<td>-0.326</td>
<td>0.0000</td>
</tr>
</tbody>
</table>

Notes: Model Summary $R^2$=0.375, adjusted $R^2$=0.361; $P < 0.0005$. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.
5.2.5.4 Stepwise Regression Particle Size >32µm <64µm

The results of the stepwise model for particle size class >32µm <64µm (dependent variable) as a function of functional diversity (plant functional types) revealed that PFT4 (Carpinus) was most correlated to this particle size fraction with adjustments for PFT 1 (Pinus and Tsuga) and PFT 2 (Poaceae) (Figures 5.14-5.15, Table 5.6). A weak but significant negative correlation was found between PFT 4 and particle size class >32µm <64µm (Figure 5.14). The analysis was repeated using all of the individual pollen taxa/species as the independent variables and revealed that Euphorbiaceae pollen had a weak but significant correlation with this particle size category explaining approximately 35% of the variability in comparison to the ~20% variability explained by PFTs 1 and 4 (Figure 5.16, Table 5.7).

Figure 5.14 Partial regression leverage plot for particle size class >32µm <64µm (dependent variable) as a function of Carpinus pollen (PFT4). The best model for the data was achieved by adding the independent variables PFT1 and PFT2.

Figure 5.15 Partial regression leverage plot for particle size class >32µm <64µm (dependent variable) as a function of PFT 2. The best model for the data was achieved by adding the independent variables PFT1 and PFT4.

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Table 5.6 Multiple regression analysis of particle size class >32µm <64µm (dependent variable) as a function of functional diversity (plant functional types). The co-dependent variables are highlighted in bold.

<table>
<thead>
<tr>
<th>Regression Variables</th>
<th>Partial Regression Coefficient</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>PFT1 (<em>Pinus, Tsuga</em>)</td>
<td>-0.186</td>
<td>0.0072</td>
</tr>
<tr>
<td>PFT2 (Poaceae)</td>
<td><strong>0.202</strong></td>
<td>0.0191</td>
</tr>
<tr>
<td>PFT3 (Cyperaceae)</td>
<td>-0.109</td>
<td>0.2933</td>
</tr>
<tr>
<td>PFT4 (<em>Carpinus</em>)</td>
<td><strong>-0.223</strong></td>
<td>0.0009</td>
</tr>
<tr>
<td>PFT5 (<em>Betula, Rosaceae</em>)</td>
<td>0.004</td>
<td>0.9807</td>
</tr>
<tr>
<td>PFT6 (<em>Thalictrum</em>)</td>
<td>0.002</td>
<td>0.9739</td>
</tr>
<tr>
<td>PFT7 (<em>Artemisia</em>)</td>
<td>0.049</td>
<td>0.5458</td>
</tr>
<tr>
<td>PFT8 (<em>Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae</em>)</td>
<td>0.097</td>
<td>0.3765</td>
</tr>
<tr>
<td>PFT 9 (Ericaceae)</td>
<td>0.000</td>
<td>0.8945</td>
</tr>
</tbody>
</table>

Notes: Model Summary $R^2=0.139$, adjusted $R^2=0.125$; $P < 0.0001$. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.

Figure 5.16 Partial regression leverage plot for particle size class >32µm <64µm (dependent variable) as a function of Euphorbiaceae pollen in the MG1 pollen record. The best model for the data was achieved by adding the independent variables Rosaceae, *Tsuga* and *Carpinus*. 
Table 5.7 Multiple regression analysis of particle size class >32µm <64µm (dependent variable) as a function of identified plant taxa (>0.5%) in the MG1 pollen record. The co-dependent variables are highlighted in bold.

<table>
<thead>
<tr>
<th>Regression Variables</th>
<th>Partial Regression Coefficient</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Betula</td>
<td>-0.053</td>
<td>0.4132</td>
</tr>
<tr>
<td>Rosaceae</td>
<td>0.144</td>
<td>0.029</td>
</tr>
<tr>
<td>Pinus</td>
<td>-0.032</td>
<td>0.630</td>
</tr>
<tr>
<td>Tsuga</td>
<td>-0.234</td>
<td>0.000</td>
</tr>
<tr>
<td>Cyclobalanopsis</td>
<td>0.126</td>
<td>0.051</td>
</tr>
<tr>
<td>Taxodiaceae</td>
<td>0.011</td>
<td>0.859</td>
</tr>
<tr>
<td>Carpinus</td>
<td>-0.164</td>
<td>0.004</td>
</tr>
<tr>
<td>Quercus</td>
<td>0.075</td>
<td>0.160</td>
</tr>
<tr>
<td>Lithocarpus</td>
<td>-0.026</td>
<td>0.635</td>
</tr>
<tr>
<td>Cyperaceae</td>
<td>-0.048</td>
<td>0.402</td>
</tr>
<tr>
<td>Poaceae</td>
<td>-0.033</td>
<td>0.606</td>
</tr>
<tr>
<td>Artemisia</td>
<td>0.010</td>
<td>0.885</td>
</tr>
<tr>
<td>Thalictrum</td>
<td>-0.009</td>
<td>0.878</td>
</tr>
<tr>
<td>Ericaceae</td>
<td>0.074</td>
<td>0.178</td>
</tr>
<tr>
<td>Euphorbiaceae</td>
<td><strong>0.347</strong></td>
<td>0.000</td>
</tr>
</tbody>
</table>

Notes: Model Summary $R^2=0.327$, adjusted $R^2=0.313$; $P<0.0001$. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.

5.2.5.5 Stepwise Regression Particle Size >64µm

The results of the stepwise model for particle size class >64µm (dependent variable) as a function of functional diversity (plant functional types) revealed that PFT1 (Pinus and Tsuga) was most correlated to this particle size fraction with adjustments for PFT 3 (Cyperaceae) (Figures 5.17- 5.18, Table 5.8). A weak but significant negative correlation was found between PFT 1 and particle size class >64µm. The analysis was repeated using all of the individual pollen taxa/species as the independent variables and revealed that Euphorbiaceae pollen had a weak but significant correlation with this particle size category explaining approximately 36% of the variability in comparison to the 32% explained by PFT1 (Table 5.9).
Figure 5.17 Partial regression leverage plot for particle size class >64µm (dependent variable) as a function of *Pinus* and *Tsuga* (PFT1). The best model for the data was achieved by adding the independent variable PFT3 (Cyperaceae).

Table 5.8 Multiple regression analysis of particle size class >64µm (dependent variable) as a function of functional diversity (plant functional types). The co-dependent variables are highlighted in bold.

<table>
<thead>
<tr>
<th>Regression Variables</th>
<th>Partial Regression Coefficient</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>PFT1 (Pinus, Tsuga)</td>
<td>-0.317</td>
<td>0.0000</td>
</tr>
<tr>
<td>PFT2 (Poaceae)</td>
<td>0.144</td>
<td>0.0511</td>
</tr>
<tr>
<td>PFT3 (Cyperaceae)</td>
<td>0.175</td>
<td>0.0106</td>
</tr>
<tr>
<td>PFT4 (Carpinus)</td>
<td>-0.053</td>
<td>0.4409</td>
</tr>
<tr>
<td>PFT5 (Betula, Rosaceae)</td>
<td>0.112</td>
<td>0.1705</td>
</tr>
<tr>
<td>PFT6 (Thalictrum)</td>
<td>-0.011</td>
<td>0.8793</td>
</tr>
<tr>
<td>PFT7 (Artemisia)</td>
<td>-0.005</td>
<td>0.9546</td>
</tr>
<tr>
<td>PFT8 (Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae)</td>
<td>-0.024</td>
<td>0.7422</td>
</tr>
<tr>
<td>PFT 9 (Ericaceae)</td>
<td>0.033</td>
<td>0.6468</td>
</tr>
</tbody>
</table>

Notes: Model Summary $R^2=0.145$, adjusted $R^2=0.136$; $P < 0.0001$. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.
Figure 5.18 Partial regression leverage plot for particle size class >64µm (dependent variable) as a function of Euphorbiaceae pollen in the MG1 record. The best model for the data was achieved by adding the independent variables Tsuga, Quercus & Cyperaceae.

Table 3 Table 5.9 Multiple regression analysis of particle size class >64µm (dependent variable) as a function of plant taxa

<table>
<thead>
<tr>
<th>Regression Variables</th>
<th>Partial Regression Coefficient</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Betula</td>
<td>0.076</td>
<td>0.288</td>
</tr>
<tr>
<td>Rosaceae</td>
<td>-0.071</td>
<td>0.338</td>
</tr>
<tr>
<td>Pinus</td>
<td>-0.122</td>
<td>0.120</td>
</tr>
<tr>
<td><strong>Tsuga</strong></td>
<td>-0.269</td>
<td>0.000</td>
</tr>
<tr>
<td>Cyclobalanopsis</td>
<td>0.029</td>
<td>0.651</td>
</tr>
<tr>
<td>Taxodiaceae</td>
<td>-0.041</td>
<td>0.521</td>
</tr>
<tr>
<td>Carpinus</td>
<td>-0.075</td>
<td>0.232</td>
</tr>
<tr>
<td><strong>Quercus</strong></td>
<td>0.153</td>
<td>0.015</td>
</tr>
<tr>
<td>Lithocarpus</td>
<td>0.024</td>
<td>0.705</td>
</tr>
<tr>
<td>Cyperaceae</td>
<td>0.192</td>
<td>0.004</td>
</tr>
<tr>
<td>Poaceae</td>
<td>0.024</td>
<td>0.727</td>
</tr>
<tr>
<td>Artemisia</td>
<td>-0.047</td>
<td>0.566</td>
</tr>
<tr>
<td>Thalictrum</td>
<td>-0.024</td>
<td>0.720</td>
</tr>
<tr>
<td>Ericaceae</td>
<td>0.041</td>
<td>0.530</td>
</tr>
<tr>
<td>Euphorbiaceae</td>
<td><strong>0.362</strong></td>
<td><strong>0.000</strong></td>
</tr>
</tbody>
</table>

Notes: Model Summary $R^2=0.302$, adjusted $R^2=0.287$, $P < 0.0001$. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.
5.2.5.6 A summary of the stepwise regression results

The results of the regression analysis indicate that:

- As PFT3 (Cyperaceae) increases, PS <4µm increases
- As PFT 2 (Poaceae) increases, PS >4µm < 16 decreases
- As PFT 3 (Cyperaceae) increases, PS >16µm <32 µm decreases
- As PFT 4 (Carpinus) increases, PS >32µm <64µm decreases
- As absolute Euphorbiaceae pollen abundances increases, PS >32µm <64µm increases
- As PFT 1 (Pinus & Tsuga) increases, PS >64µm decreases.
- As absolute Euphorbiaceae pollen abundances increases, PS >64µm increases

5.2.6 A comparative analysis of the statistical results

The preliminary testing of the data using Spearman Rank correlation coefficients revealed weak but significant correlations between climate and the PFT groupings 1 (Pinus and Tsuga), 3 (Cyperaceae) and 5 (Betula and Rosaceae) i.e. climate is positively correlated with an array of plant functional diversity types in the lake Mugucuo catchment during the Holocene. Interestingly, the climate record was only found to be significantly (albeit weakly) correlated with the particle size fraction >16µm <32 µm, thereby implying that climate is not generally significantly correlated with lake catchment dynamics at Lake Mugucuo during the Holocene. Analysis of the plant functional diversity data and particle size data revealed weak but significant correlations between PS <4µm and PFT3 (Cyperaceae), PS >4µm < 16µm and PFT1 (Pinus and Tsuga), PS >16µm <32 µm and PFT 3 (Cyperaceae), PS >32µm <64µm and PFT1, and finally PS >64µm and PFTs 1 and 3. These preliminary results implied that lake catchment dynamics do correlate to plant functional diversity in the Lake Mugucuo catchment and therefore vegetation is likely the mediating factor between climate and landscape change.

Further investigation revealed that both plant functional diversity and lake catchment dynamics respond to changes in climate on millennial timescales. The cross-correlation analysis revealed that the lake catchment dynamics proxies PS <4µm, >4µm < 16µm, >16µm <32 µm i.e. the finer classes of sediment respond to changes in plant functional diversity on millennial timescales. Conversely the more coarse particle size fractions PS >32µm <64µm and PS >64µm were found to respond to changes in plant functional diversity on centennial timescales, in particular grass (PFT2) and shrub species (PFT7). This result seems plausible as plants shelter and fix the soil with their roots (Körner, 2002 and Gyssels et al., 2005) and they can act as a physical barrier, altering sediment flow at the soil surface (Van Dijk et al., 1996 and Lee et al., 2000) on short-term timescales ($10^2$ yr). Grasses and shrubs are effective short-term sediment trappers (Körner, 2002) which may explain the centennial-scale lags between the coarse particle size proxies (i.e. those that are indicative of increased erosion events in the record) and their relationships with grass and shrub taxa in the MG1 record.
The cross-spectral results revealed some interesting relationships between the proxies. The data showed that common periodicities are found between climate and plant functional diversity in the Mugecuo catchment. The dominant correlating periodicities were found to range from ~2,800 to ~3,400 years between climate and predominantly grass (PFT2), herb (PFT6), shrub (PFT9) and tree (PFT4) taxa i.e. a range of functional diversity types. This suggests that changes in the plant functional diversity record at Lake Mugecuo correlate to changes in climate on millennial timescales. Thereby providing evidence that climate is a main driver of changes in plant community dynamics as discussed in section 1.2.1. The cross-spectral analysis also showed that common periodicities were found between the climate data and the particle size fractions >4µm < 16µm and PS >64µm of 2,800 years. However, the particle size fractions <4µm, >16µm <32 and >32µm <64µm only displayed correlations with climate on timescales >10,000 years. In a data sequence of ~12,500 years it is highly unlikely that these trends of >10,000 years are real due to the length of the timescale of the dataset being investigated i.e. these are not treated as realistic trends here. However, the data does indicate that changes in climate are in some way related to changes in sediment size in the catchment, but something else is likely driving changes in particle size variability on shorter-term timescales (<2,800 years).

Further analysis of the data indicated that correlating periodicities were found between PS <4µm and PFT 4 and PFT8 which are predominantly representative of tree taxa in the MG1 record i.e. when the catchment has a substantial tree population there tends to a dominance of fine sediment entering the lake. The data also revealed that correlating periodicities were found PS >4µm < 16µm and PFT2, PFT3, PFT4 and PFT9 which are representative primarily of grass, sedge, shrub and tree taxa thereby suggesting a mixed catchment vegetation with a forest component and alpine meadow which correlates with a slightly coarser sediment size. These correlating periodicities were on timescales ranging from 1,445 to 2,413 years i.e. shorter timescales than that of the climate data versus this particle size which as mentioned above was approximately 2,800 years. The coarser particle size fraction PS >32µm <64µm was also found to have correlating periodicities with PFT 1 (tree taxa), PFT4 (tree taxa) and PFT6 (herbaceous taxa) on much shorter timescales than those found with the climate data. PS >64µm was found to be correlated to plant functional diversity (PFT 8) again on a much shorter timescale than the correlation found between the climate data and this particular particle size fraction. At this point, it was reasonable to assume that plant functional diversity has a greater correlation to particle size data (the proxy for landscape/catchment dynamics in this study) than the particle size data has to the climate data based on higher R² values (see the Spearman Rank results section 5.2.2 and Appendix A.2) i.e. there is not a significant correlation between lake catchment dynamics and changes in climate at Lake Mugecuo during the Holocene. But there is a significant correlation between changes in plant functional diversity and changes in catchment dynamics in the lake Mugecuo catchment. Interestingly, the correlating periodicities between >16µm <32 µm and all of the plant functional types were found to be on timescales >10,000 years, similar to the correlation of this particle size fraction to the climate data. Therefore it is not evident in this data as to what could be
driving changes in this particular particle size in the catchment through time as this trend may again not be a realistic one based on the timescales of the datasets.

The overall results of the cross-correlation and cross-spectral analysis support the theory that vegetation mediates the effects of climate change on lake catchment dynamics at Lake Mugecuo. The results revealed that on the whole, plant functional diversity is more correlated to lake catchment dynamics on shorter timescales than climate is to lake catchment dynamics. The next phase of analysis therefore aimed to infer a causal relationship between plant functional diversity to lake catchment dynamics in the MG1 record. The stepwise regression analysis revealed some notable results. The analysis showed that Cyperaceae was the most important variable explaining changes in the particle size fraction PS <4µm ($r^2$=0.29), but only when Poaceae, Betula, Rosaceae and Ericaceae were included in the model. This suggests that one species alone does not influence this size of sediment in the catchment but that an array of plant functional diversity types influence catchment dynamics at this scale/level i.e. a mixed vegetation of sedges (Cyperaceae) in conjunction with grass (Poaceae), shrub (Rosaceae, Ericaceae) and tree taxa (Betula) influence this sediment size. In a study of vegetative filter strips (VFS) by Munoz-Carpena et al., (1999) showed that vegetation characteristics, including types, density and height influence the ability of vegetation to trap sediment on a hillslope. It has also been shown that members of the Cyperaceae family (e.g. Carex) are effective at trapping sediment upslope of the plant (Isselin-Nondedeu and Bédécarrats, 2007). This would suggest that when there is a high abundance of sedges along with grasses and shrubs, which previously mentioned are effective short-term sediment trappers (Körner, 2002), coarser sediment does not get easily transported into the lake whilst finer sediment does.

The analysis revealed that particle size fraction >4µm < 16µm was most highly correlated with Poaceae (grasses). However, they were negatively correlated and therefore it was inferred that as Poaceae pollen increases in the MG1 record, PS >4µm < 16µm declines in the lake record. The stepwise regression model indicated that Poaceae was most significantly correlated with PS >4µm < 16µm ($R^2$) when found with Pinus, Tsuga, Carpinus and Ericaceae. It is important to highlight at this point that these taxa were not negatively correlated to this particle size which suggests that an increase in Poaceae and a decrease in these other taxa result lower levels of PS >4µm < 16µm from entering the lake. Figure 5.12 indicates that this is a very weak trend and the cross-spectral analysis revealed that Poaceae was correlated with this particle size on timescales of approximately 1,580 years. The results therefore indicate that changes in PS >4µm < 16µm are correlated to changes in Poaceae in the MG1 record on millennial timescales. Another study of vegetative filter strips (VFS) by Pan et al. (2010) which involved experiments on grass strips retarding severely eroded steep cropland runoff under simulated rainfall revealed that grass strips were more effective in trapping sediments coarser than 10 µm but <25 µm. Pan et al. (2010) demonstrated that these grass strips could remove more than 40% of sediment from upslope and that they are highly effective in trapping coarse sediment of that size which explains the negative correlation of Poaceae with PS >4µm < 16µm in this study.
Cyperaceae (PFT3) was found to be the most correlated variable with PS >16µm <32 µm when adjusted for Carpinus (PFT4), Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae (PFT8) and Ericaceae (PFT9), but this was a negative correlation thereby suggesting that lake at this spatial scale an increased presence of Cyperaceae in the catchment results in declining levels of PS >16µm <32 from entering the lake. However, the cross-spectral analysis results indicate that this relationship is likely on a timescale of >10,000 years, therefore it does not exactly tie in with the aim of assessing the relationship between catchment dynamics and plant functional diversity on shorter-term timescales and due to the nature of the timescales these trends are not treated as realistic here.

Both PFT 4 (Carpinus) and PFT2 (Poaceae) were the most correlated plant functional diversity variables for PS >32µm <64µm. PFT 4 was negatively correlated to PS >32µm <64µm and these two variables were found to be correlated on timescales of approximately 1,408 years in the cross-spectral analysis. On the other hand, PFT (Poaceae) was positively correlated to PS >32µm <64µm and these two signals were found to be correlated on timescales of approximately 2,413 years. Figures 5.12 and 5.13 indicate that on the whole these are very weak trends between PS >32µm <64 and PFTs 4 and 2. In testing all of the pollen taxa individually against the particle size data it was found that Euphorbiaceae (classified as a shrub in this study) on its own was the most correlated variable with PS >32µm <64µm explaining 34.7% of the data. When combined with the Rosaceae, Tsuga and Carpinus pollen data i.e. a mixture of shrub and tree taxa, it appears that an array of plant functional types co-exist together to influence lake catchment dynamics on this spatial scale. Overall the data indicates that when shrubs and grasses increase in the catchment and tree abundances decrease, more coarse sediment enters the lake. It is likely here that changes in the tree and shrub component of catchment vegetation result in a greater sediment supply of coarse material to the lake when there is a lesser presence of sedge and grass taxa which have been found to be effective trappers of coarse sediment in the Lake Mugecuo (discussed in the previous paragraph). Trees help to stabilize soils against erosion and reduce soil disturbance (Burylo et al., 2011). It is possible that a change in the forest component by tree turnover events may induce increased catchment erosion (Roering et al., 2010), which could result in increased coarse material entering the lake. The relationship between changes in plant community dynamics and the erosion record for lake Mugecuo based on $^{10}$Be measurements is further analysed in the following sections.

Cross-spectral analysis of the Euphorbiaceae data against PS >32µm <64µm (Figure 5.18) revealed that catchment dynamics on this scale respond to variability in Euphorbiaceae on timescales of c. 1,689 years. Euphorbiaceae on its own was also the most correlated variable with PS >64µm explaining 36.2% of the data in this case. It was however, closely followed by PFT1 (Pinus and Tsuga) which were negatively correlated with the particle size fraction (31.7%). Once again the data suggest that changes an increase in shrub vegetation and a decrease in tree taxa result in increased coarse sediment size presences in the lake record.
Figure 5.19 Cross-spectral analysis of the Euphorbiaceae data against PS >32µm <64µm. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.

5.3 Cosmogenic $^{10}$Be erosion rates

The steady erosion rates based on the $^{10}$Be concentrations measured in bedrock, fluvial sediment samples and lacustrine palaeo-deposits of quartz from the Lake Mugecuo catchment are discussed here in relation to the MG1 record and the catchment dynamics of the lake over the last 3,800 years. By establishing a unified core chronology as outlined in section 2.4.3, erosion-rates measured in the shallow cores could be related equally to the pollen, organic, and lithological proxies in the deep water core as each of the cores spanned concurrent periods. Figure 5.20 thus highlights the position of the calculated palaeo-erosion rates (outlined in section 3.5.3) in relation to the MG1 stratigraphic record in order to illustrate how these measurements relate to the wider multi-proxy record for the MG1 core. The palaeo-fluvial, modern fluvial and bedrock samples in this study provide basin-scale rates of denudation. As previously mentioned in Chapter 1, the term 'fluvial filtering' has been proposed to refer to the role of the cascade of biotic and sediment storages located along river systems in trapping nutrients that might otherwise reach a lake bed (Meybeck and Vorosmarty 2005). Such filtering can result in a ‘smoothing’ of the signal of an individual proxy and fluvial filtering processes have the power to greatly influence lake sediment that is archived in a proxy sequence. Unlike the lake proxy records discussed above which have been shown to be sensitive to filtering by vegetation such as sedges and grasses, the $^{10}$Be measurements give a true picture of catchment dynamics as they are insensitive to changes in sediment storage (Granger et al 1996). $^{10}$Be erosion rate measurements give catchment wide erosion rates for specific periods of time that are unrelated to depositional environment and they are unaffected by fluvial filtering because they are average results integrated over 100s years (meaning they are less likely affected by sediment trapping) and therefore provide an independent test of what these other proxies may tell us.
5.3.1 Integration times and erosion rates

The time-scales over which denudation rates integrate is equivalent to the time required to remove about 60 cm of bedrock (Granger et al., 1996; von Blanckenburg, 2006) and is thus inversely correlated to the denudation rate (Figure 5.21). Therefore, erosion rates are only valid over the time required to erode 0.6mm depth of material e.g. if erosion rate is 1cm/yr the integration time will be 60 years.

Figure 5.20 Stratigraphic record of the MG1 core (now unified with MG3 & MG4) with calibrated chronology attached and calculated palaeo- and modern erosion rates (mm/yr).
Figure 5.21 Cosmic ray neutrons, the primary producers of most cosmogenic nuclides including $^{10}$Be, are attenuated rapidly below Earth’s surface as a function of the mass penetrated. (A) Attenuation in rock ($\rho = 2.7 \text{ g cm}^{-3}$). (B) Attenuation in soil ($\rho = 1.2 \text{ g cm}^{-3}$) (Bierman & Nichols, 2004).

Integration times for the palaeo-erosion rates were calculated here by dividing the maximum erosion rate into 600mm (i.e. 60cm). The integration times for the Lake Mugecuo palaeo-erosion sediment samples range from ~1,146 to ~3221 years. These integration times are outlined in Table 5.10 and Figure 5.22 in relation to the MG1 record and the time-span each erosion rate integrates over. The integration times for the modern erosion rates range from ~1,875 to ~15,000 years. In terms of interpreting the overlaps, further modelling of the data would be required such that the effects of variable conditions in erosion and overlying soil depth could be evaluated (e.g. Reinhardt et al., 2005; Heimsath, 2006). However, such modeling goes beyond the scope of this project and was not pursued here.
Table 5.10 Integration times for each of the palaeo-, modern fluvial and bedrock erosion rates for the Lake Mugecuo catchment.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Midpoint Sample Age kcal.yr BP</th>
<th>Midpoint Sample Depth Cm</th>
<th>Maximum Erosion Rate (mm/yr)</th>
<th>Erosion Rate Error Total (mm/yr)</th>
<th>Calculated Integration Times (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MG3a</td>
<td>3.0</td>
<td>112</td>
<td>0.21</td>
<td>0.02</td>
<td>2807</td>
</tr>
<tr>
<td>MG3b</td>
<td>3.4</td>
<td>120</td>
<td>0.22</td>
<td>0.02</td>
<td>2760</td>
</tr>
<tr>
<td>MG3c</td>
<td>3.8</td>
<td>130</td>
<td>0.52</td>
<td>0.05</td>
<td>1146</td>
</tr>
<tr>
<td>MG4a</td>
<td>0.7</td>
<td>21</td>
<td>0.19</td>
<td>0.02</td>
<td>3221</td>
</tr>
<tr>
<td>MG4b</td>
<td>1.2</td>
<td>35</td>
<td>0.20</td>
<td>0.02</td>
<td>3020</td>
</tr>
<tr>
<td>MGDR</td>
<td>Modern</td>
<td>-</td>
<td>0.17</td>
<td>0.02</td>
<td>3,529</td>
</tr>
<tr>
<td>MGMR</td>
<td>Modern</td>
<td>-</td>
<td>0.32</td>
<td>0.04</td>
<td>1,875</td>
</tr>
<tr>
<td>MGCR</td>
<td>Modern</td>
<td>-</td>
<td>0.04</td>
<td>0.01</td>
<td>15,000</td>
</tr>
</tbody>
</table>

Figure 5.22 MG1 core stratigraphy with integration times for each measured palaeo-, fluvial and bedrock erosion rate measured at the Lake Mugecuo catchment.
The samples MG3a & MG3b were shown to have approximately equal steady erosion rates of 0.21 ± 0.02 mm/yr and 0.22 ± 0.02 mm/yr respectively. The younger of the sediments measured (MG4a & MG4b) were also shown to have relatively similar steady erosion rates of 0.19 ± 0.02 mm/yr (MG4a) and 0.20 ± 0.02 mm/yr (MG4b). On the whole, these measurements provide evidence of a relatively stable catchment. However, a notable increase in erosion rates at sample MG3c indicates a significant increase in erosion rates to 0.52 ± 0.02 mm/yr at approximately 3.8 kcal.yr BP which is an indication of a marked shift in catchment-wide denudation conditions during this period.

The modern catchment erosion rates calculated from fluvial sampling site A along with the bedrock sample revealed steady erosion rates of approximately 0.17 ± 0.02 mm/yr and 0.04 ± 0.01 mm/yr, respectively. A notable difference was found in the modern catchment rate calculated from fluvial sampling site B which gave steady erosion rates of 0.32 ± 0.04 mm/yr. Based on the topography at this sampling site, the material taken from this part of the lake is coming from a different source area (see Figure 5.23). The material represented by the greater erosion rate is coming from a much steeper terrain (Figure 5.24) with higher rates of erosion (c.f. Kirchner et al., 2001). Because all erosion is fundamentally driven by gravity, steeper slopes and greater relief (Figure 5.25) result in increased erosion rates (Montgomery and Brandon 2002). Referring back to Figure 2.3, these differences in 10Be concentrations also provide direct evidence of the direction of movement of sediment through the lake catchment. The bedrock sample has a higher 10Be concentration (i.e. a lower erosion rate) as it has had a greater exposure time to cosmic rays in comparison to the other two modern samples due to their positions in the catchment and the pathway of sediment as it transported through the catchment system (see Figure 2.3). The differences in the modern 10Be-based erosion rate estimates in the Mugecuo catchment therefore indicate a high degree of spatial variation in rates of erosion in this high-relief catchment.
Figure 5.23 Sampling sites of bedrock, fluvial, and palaeo-deposit material for $^{10}$Be concentration measurement in the Lake Mugecuo catchment. A Google earth image shows the likely direction of flow of material into the lake at each sampling site.
Figure 5.24 The differences in slope of the two modern fluvial sampling sites (red = high) green = low. The Slope tool in ArcGis calculates the maximum rate of change between each cell and its neighbouring cells. The lower the slope value, the flatter the terrain; the higher the slope value, the steeper the terrain.

Figure 5.25 Plot of long-term erosion rates versus mean slope for a transect across the Olympic Mountains in a study by Montgomery and Brandon (2002).

There still remains a need to explain the single high erosion rate recorded at 3.8 kcal.yr BP in the Mugecuo record. Large bedrock landslides can incise to depths greater than the attenuation length of cosmic rays, thus mobilizing sediments with little or no cosmogenic nuclide abundance which will
dilute the $^{10}$Be concentration in the fluvial sediment, yielding apparently higher erosion rates (Niemi et al., 2005). It is most likely that the high erosion rate recorded at approximately 3.8 kcal.yr BP was therefore caused by a large-scale localized landslide. Following the methods outlined in Reinhardt et al. (2007) which addresses the issues of interpreting erosion rates measured in rapidly eroding terrain, theoretical analysis of the data implies that the palaeo-erosion rate estimates are generally unaffected by frequent shallow (<3m) regolith or bedrock landslides due to internally consistent rates measured in the part of the catchment which samples MG4a, MG4b, MG3a, MG3b and fluvial sampling site A were retrieved. However, the difference in the bedrock erosion rate and the erosion rates calculated from the two fluvial sampling sites would suggest that the catchment is not in equilibrium as the bedrock samples (n=10) are eroding at a slower rate. Disequilibrium landforms are those that tend toward equilibrium but have not had sufficient time to reach such a condition (Renwick, 1992). The integration time of the modern bedrock sample is approximately 15,000 years, therefore the faster erosion rates measured in the palaeo- and modern fluvial samples essentially indicate that erosion has accelerated during the Holocene in the Mugecuo catchment. These results are further discussed in section 5.4.2-3 in order to determine the potential causes of the differences in erosion rates for the Mugecuo catchment over the last 3,400 years.

5.4 Palaeo-erosion rates and the Lake Mugecuo catchment dynamics proxies

The relationship between the measured palaeo-erosion rates (mm/yr) for the last 3,800 years from the Lake Mugecuo record and the pollen, plant functional diversity proxies and lake catchment dynamics proxies from the MG1 record are discussed here. The discussion of the proxies is divided into two sections based on the zoning of the data previously discussed in sections 4.2 and 5.1.

5.4.1 MP-Z 4 (5.8-2.3 kcal.yr BP; 172-85cm)

The measured palaeo-erosion rates spanning the last 3,800 years in the Lake Mugecuo record are plotted against the organic and lithological proxies of the Mg1 record in Figure 5.26. Although sedimentation levels remain low during the time span that the palaeo-erosion rate samples cover in this zone, a marked increase in erosion rates at sample MG3c indicates a distinct difference in minimum erosion rates at approximately 3.8 kcal.yr BP in comparison to the erosion rates calculated at 3.0 kcal.yr BP (MG3b) and 3.4 kcal.yr BP (MG3a). It is most likely that a denudation event i.e. the localized effects of a large landslide explains the increased erosion rate at approximately 3.8 kcal.yr BP. In a catchment where a deep landslide has recently occurred, the addition of nuclide-poor landslide detritus to the fluvial system will dilute the $^{10}$Be concentration in the fluvial sediment, yielding apparently higher erosion rates (Niemi et al., 2005). Therefore, a landslide during this period may have resulted in a catchment-wide denudation event which led to a pulse of sediment with a greater erosion rate of $0.52 \pm 0.02$ mm/yr entering the lake. An important point to raise at this point is the lack of any significant evidence of a major change in erosion rates around the time of the 3.8 kcal.yr BP event in the organic and lithological proxies. This finding suggests that particle size data and the other
geochemical and organic proxies are in fact not useful for assessing lake catchment dynamics in palaeoenvironmental reconstructions and thus changes in erosion in a catchment over time.

The palaeo-erosion rates 3a, 3b and 3c are plotted against the pollen percentages of arboreal (AP) and non-arboreal (NAP) taxa along with summary TPC (total pollen concentration) total AP and total NAP percentages in Figure 5.27. The increased erosion rates at approximately 3.8 kcal.yr BP coincides with a notable drop in Cyperaceae and an overall increasing trend in arboreal pollen percentage levels. It has been discussed that members of the Cyperaceae family (PFT3) are effective at trapping sediment upslope of the plant (Isselin-Nondedeu and Bédécarrats, 2007). In tandem with this high magnitude denudation event likely caused by landsliding activity, the decline in Cyperaceae at this point may have resulted in less sediment trapping by the edges of the lake as sedge species are generally found in marshy habitats which allowed for the transport of landslide detritus into the lake. The plant functional diversity indicators (Figure 5.28) reveal a change in plant community dynamics, with an increase in PFT 8 pollen percentages (Quercus, Taxodiaceae, Cyclobalanopsis, Lithocarpus, Euphorbiaceae) and a slight increase in dominant PFT 1 (Pinus and Tsuga) during this increased period of catchment denudation. PFT 8 is the main contributor to the record with levels of up to 30%. This change in the pollen record likely signifies a change in the dynamics of the forest component of the catchment which in conjunction with a declining trend in Cyperaceae (PFT3) taxa likely coincided with a mass-denudation event (i.e. landslide) in the catchment reflected in the measured palaeo-erosion rate of 0.52 ± 0.02 mm/yr at 3.8 kcal.yr BP. Vegetation protects a landscape from erosion by reducing water erosion (Rey, 2003 and Puigdefabregas, 2005) and increasing the infiltration rate in soil (Graeme and Dunkerley, 1993, Ziegler and Giambelluca, 1998 and Wainwright et al., 2002). Other graminoids (such as Poaceae) have also been shown to form mounds and trap sediment (Isselin-Nondedeu and Bédécarrats, 2007). A general decline in shrubs and grasses from approximately 4.0 kcal.yr BP may also have contributed to increased potential for landsliding conditions in the catchment as shrubs generally act by stabilizing the soil and acting as physical barriers, thus, reducing transport of soil particles down slope.
Figure 5.26 Organic and lithological proxies of zone MPZ-5 (5.8-2.3 kcal.yr BP; 172-85 cm) of the MG1 record with calculated erosion rates (mm/yr).
Figure 5.27 MP-6 pollen percentages of arboreal (AP) and non-arboreal (NAP) taxa including trees, shrubs, herbs, grasses and sedges of zone P-5 (5.8-2.3 kcal yr BP; 172-85 cm). Spores/ferns and aquatics are also included. Due to the large number of species identified, taxa with an abundance <0.1% are excluded from this diagram. Calculated erosion rates (mm/yr) are marked for reference.
5.4.2 MPZ- 5 (2.3-0.0 kcal.yr BP; 85-0cm)

The measured palaeo-erosion rates spanning the last 1,200 years in the Lake Mugecuo record are plotted against the organic and lithological proxies of the Mg1 record in Figure 5.29. The palaeo-erosion rates 4a and 4b indicate that during this time-span (0.7-1.9 kcal.yr BP), no major denudation events occurred in the Lake Mugecuo catchment as the rates are the same (0.19-0.20 mm/yr). There is also very little evidence in the organic proxies during this period of any major erosion/deposition events or in-wash of sediment into the lake (Figure 5.28). However, the particle size data indicates increased levels of sand from approximately 1.8-1.2 kcal.yr BP. This is not reflected in the palaeo-erosion rate data where there are extremely low levels of sand (0% from approximately 1.0-0.8 kcal.yr BP), yet the calculated erosion rates during these two periods are the same. Sedimentation rates between dated points also indicate a slight difference in deposition rates during the two periods covered by these two palaeo-erosion rates. Inconclusive findings such as this highlight the benefit of adopting a multi-proxy approach into the investigation of catchment dynamics as the palaeo-erosion rate samples taken from the mouth of the river delta indicate that no major denudation changes are
occurring in the catchment during this period i.e. increased levels of sand in the particle size data do not necessarily indicate increased erosion in the catchment. Another interesting point is that the study by Zhou et al. (2001) of a trench on the Southwestern bank of the Mugecuo lake indicated that two earthquakes occurred at the Lake Mugecuo catchment at approximately 1.2 yr BP and 1.6 yr BP. There is very little evidence of earthquakes in the organic proxies during these two periods but the 1.6 yr BP earthquake coincides with the rise in sand % in the lake record. However, taking the integration times of the two $^{10}$Be erosion rate samples into consideration there is no evidence of this type of measured erosion rates which span that time period. This finding also supports the idea that particle size data is in fact not a useful proxy for lake catchment dynamics in palaeoenvironmental reconstructions.

The palaeo-erosion rates are plotted against the pollen percentages of arboreal (AP) and non-arboreal (NAP) taxa along with summary TPC (total pollen concentration) total AP and total NAP percentages for MPZ-5 (Figure 5.31). Very little change occurs in the MG1 pollen record during this period which supports constant palaeo-erosion rates at 0.7 and 1.2kcal.yr BP. When plotted against the plant functional diversity data (Figure 5.30), a notable change in PFT 1 ($Pinus$ and $Tsuga$) occurs at approximately 1.3 kcal.yr BP which is also indicated by a drop in total AP% in Figure 5.31. This change indicates that the decline of tree taxa in the catchment during this period did not have a major impact in the catchment dynamics i.e. it did not result in a major erosion/denudation event or influx of material into the lake. However, the relatively high presence of non-arboreal pollen in conjunction with relatively high contributions from Cyperaceae pollen at this point may have contributed to increased sediment trapping (c.f. Isselin-Nondedeu and Bédécarrats, 2007) and/or soil stabilization in the catchment which prevented a major denudation event from occurring in the catchment during this period contrary to the conditions seen during the 3.8 kcal.yr BP changes in erosion rates.

These results and the results of the statistical analyses outlined in section 5.2 are drawn together and further discussed in Chapter 6.
Figure 5.29 Organic and lithological proxies of zone MPZ-6 (2.30 kcal.yr BP; 85-0cm) of the MG1 record with calculated erosion rates (mm/yr).

Figure 5.30 Plant functional diversity proxies (PFTs 1-9) of MPZ-6 (2.30 kcal.yr BP; 85-0cm) of the MG1 record with calculated erosion rates (mm/yr).
Figure 5.31 MG1 Pollen percentages of arboreal (AP) and non-arboreal (NAP) taxa including trees, shrubs, herbs, grasses and sedges of zone P-6 (2.3-0 kcal yr BP; 85-00cm). Spores/ferns and aquatics are also included. Due to the large number of species identified, taxa with an abundance <0.1% are excluded from this diagram. Calculated erosion rates (mm/yr) are marked for reference.
6) Chapter 6 A multi-proxy reconstruction at Lake Mugecuo - a critical analysis

The Lake Mugecuo record provides an environmental, vegetation and climatic history of events from the end of the Late-Glacial period to the Late Holocene c. 12.6-0.0 kcal.yr BP at a remote alpine lake catchment in Sichuan Province, Southwest China. The main aim of this study was to reconstruct past catchment vegetation dynamics at Lake Mugecuo and past variation in sediment dynamics within the lake in order to determine a) how a dynamic high alpine lake catchment system in Sichuan Province, Southwestern China responded to major long-term changes in climatic change i.e. changes in the East Asian Summer Monsoon since the Early Holocene? and b) to determine the effects of plant functional diversity on catchment dynamics through time in order to assess whether plant functional diversity mediates the effects of climate on lake catchment dynamics. The results of each of the proxies used in this study are drawn together and discussed in this section and a critical analysis of the multi-proxy approach adopted in this study is also discussed. The applicability and validity of this approach and its findings to the wider Sichuan region and similar catchment systems are also outlined.

6.1 Integrated discussion of multi-proxy results

The deep water core MG1 formed the focus of the palaeoecological and the majority of the palaeolimnological work of this project while the shallow water cores MG3 and MG4 provided evidence of palaeo-erosion rates for the catchment during the late Holocene period. Figure 6.1 presents an amalgamation of the data analysed in Chapter 5, while Table 6.1 summarises the main environmental, vegetation, and climatic characteristics of each multi-proxy zone inferred from the Lake Mugecuo proxies. The findings and implications of these analyses are further discussed in this section.
Figure 6.1 Summary of the plant functional diversity, particle size, smoothed δ¹⁸O speleothem records and palaeo-erosion rates which formed the focus of the quantitative analysis of the Lake Mugecuo record in order to test the main research hypotheses.
Table 6.1 Summary of the environmental, vegetation and climatic characteristics of each multi-proxy zone inferred from the Lake Mugecuo proxies.

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The MG1 record indicates that since the Younger Dryas through to the late Holocene (ca. 12.6-0.0 kcal.yr BP) Lake Mugecuo was a relatively stable lake characterised by periods of enhanced activity on millennial and sub-millennial timescales. The record indicates that physical weathering was dominant in the catchment throughout the Holocene with evidence of periods of cold and dry weathering conditions (refer to section 5.1). The Holocene history of the human occupation of this region is poorly known. However, no traces of early human activities which may have influenced the hydrological regime of the palaeo-lake has been found in the Lake Mugecuo catchment. Sedimentation levels remain relatively stable with little evidence of increased erosion due to human activity in the area i.e. slash and burn agriculture which is a popular method of forest removal in this region (Liu & Liu, 1985).

The D4 δ^{18}O speleothem record, a well-dated high-resolution record of the changes in the East Asian Summer Monsoon (EASM) during the Holocene (Dykoski et al., 2005), indicates a weakening EASM intensity during the Holocene, which is evidenced by changes in the composition of the vegetation at Lake Mugecuo (Chapter 4). The MG1 pollen record indicates that the Late Glacial period (with some evidence of Younger Dryas conditions) at Lake Mugecuo (12.6-11.9 kcal.yr BP) was characterised by cold and wet winter conditions while the Early Holocene (approximately 11.9-8.3 kcal.yr BP) was characterised by cold temperatures and dry winter conditions. The mid-Holocene period at Lake Mugecuo (approximately 8.3-2.3 kcal.yr BP) was characterised by distinct wet and dry seasons with a strong Asian summer monsoon prevailing. Conversely, the Late Holocene (approximately 2.3 to 0.0 kcal.yr BP) at Lake Mugecuo was characterised by a shift to warmer temperatures with less rainfall. On the whole, the seasonal and spatial patterns changes implied by the Lake Mugecuo reconstruction are consistent with previous vegetation-based estimates of Holocene climate (c.f. Shi et al., 1993; Yu et al 2000; Shen et al., 2005; Herzschuh, 2006). The main drivers of these changes at Lake Mugecuo were distinguished based on the PCA ordination results (section 4.3.1) which indicated that effective moisture followed by summer temperature were the main drivers of vegetation change at Lake Mugecuo during the Holocene.

The classification of the MG1 pollen data into plant functional types (PFTs) provided an appropriate means of assessing changes in plant community dynamics in the Lake Mugecuo catchment during the Holocene. The plant functional diversity data revealed that temperate deciduous broadleaved taxa dominated the early portion of the Lake Mugecuo record (12.5 to 8.3 kcal.yr BP). Over time, due to changes in the EASM regime, cool-temperate evergreen forest taxa began to take hold from approximately 8.3-2.3 kcal.yr BP. The late Holocene saw an increased presence of warm-temperate evergreen forest species, indicative of the weakening of the EAM intensity (Dykoski et al., 2005; Herzschuh, 2006) with a long-term shift towards increasingly drier and warmer climatic conditions as per Jarvis (1993), Shen et al. (2005; 2006). The Mugecuo record is generally characterised by relatively low levels of biological and organic productivity, with a dominance of aquatic-driven productivity in the sedimentary record (section 5.1). Over time, the Lake Mugecuo catchment has become less densely vegetated and less organically productive, suggesting a long-term shift towards
a less fluvially dynamic, productive catchment over millennial timescales. The increased levels of Cyperaceae and Poaceae in the mid-to-late-Holocene are likely associated with wetland development (c.f. van Campo et al., 1996) at Lake Mugecuo and the gradual expansion of meadow and grasslands (section 4.2) likely coincides with a slowdown of the fluvial dynamics in the catchment valley.

The palaeo-erosion rates measured in the MG3 and MG4 shallow-water cores which were linked to the MG1 record through a unified core chronology based on age-depth modeling using Oxcal v. 4.17 (Bronk Ramsey, 1995; Bronk Ramsey, 2008), provide an independent snap-shot of catchment-wide erosion rates on a very distinct temporal scale for this section of the Lake Mugecuo record. Bearing in mind that it has already been established that low-velocity flows were in operation at the lake throughout the Holocene, the MG1 record alone was unlikely to capture the majority of erosion events due to the energy required to transport coarse material to the deepest part of the lake basin where the MG1 core was extracted.

It is important to stress that cosmogenic \(^{10}\)Be erosion rates are independent of processes of sedimentation, therefore the cosmogenic erosion rates are an excellent supplementary proxy which provided added information on the catchment dynamics at Lake Mugecuo spanning the last 3,400 years.

Overall the data indicates that the ~3.8 kcal.yr BP increase in erosion rates to approximately 0.52 mm/yr aside, palaeo-erosion rates in the catchment over the last 3,800 years were relatively similar ranging between 0.19-0.22 mm/yr. These rates were also relatively consistent with the modern erosion rates (approximately 0.17 mm/yr; Fluvial sample A) calculated for the catchment. However, the modern erosion rate measured at fluvial sampling site A was quite different from the erosion rates measured at fluvial sampling site B (0.32mm/yr) which likely indicates a difference in the sediment source areas of these two sampling sites (Figures 5.23-5.24). The differences in topography for these two sampling sites indicate that fluvial sampling site B is a sink for material coming from a much steeper terrain with greater potential of shallow landsliding and debris slides (c.f. Kirchner et al., 2001). The two modern fluvial rates differed by one order of magnitude variation with regards to the bedrock sampling site (0.07mm/yr) which suggests that the catchment is actually in disequilibrium as the bedrock samples (n=10) are eroding at a slower rate (c.f. Renwick, 1992; Heimsath, 2006). This also indicates that sediment found at this location have been exposed to greater concentrations of cosmic rays and thus provides evidence of the pathway of sediment transport through the catchment system. As a result of extensive and on-going tectonic activity in the Hengduan Mountains, it is an active fault zone with many earthquakes (Meyerhoff et al. 1991). The period of higher erosion at 3.8 kcal.yr BP was most likely a result of tectonic activity which then resulted in a mass denudation event (e.g. a large-scale landslide) in the catchment as opposed to sediment trapping by vegetation which would be less likely to exert a topographic control of this magnitude. However, based on the changes in plant functional diversity (albeit very small) during the same time period as the increased rate of erosion measured at 3.8 kcal.yr BP as previously outlined in section 5.4, it is inferred that vegetation does play a part in controlling the influx of material into the lake and therefore likely ‘mediates’ the effects of such large denudation events, a concept which is discussed further below.
In Chapter 1 it was highlighted that numerous studies have shown that long-term erosion rates do not correlate with mean annual precipitation and temperature, implying that long-term erosion rates are not sensitive to catchment hydrology (von Blanckenburg, 2005). However, given that changes in precipitation and temperature can alter the ecology of any region (Stenseth et al., 2003), it is argued here that it is intuitively reasonable to suggest that climate change indirectly perturbs a landscape through its influence upon a landscapes ecosystem following the ideas outlined in Reinhardt et al. (2010). It was inferred in Chapter 4, that fluctuations in pollen diversity and plant community dynamics during the Holocene were most likely attributed to changing climatic conditions (i.e. changes in the EASM) over time. Statistical analysis of the data has shown that plant functional diversity and lake catchment dynamics respond to changes in climate on millennial timescales in the Lake Mugecuo catchment. However, interactions between plant functional diversity and lake catchment dynamics were generally found to vary on shorter timescales (<2,800yr). It is therefore argued here that the environmental shifts captured in the Lake Mugecuo record are primarily driven by changes in plant functional diversity and that vegetation likely mediates the effects of climate change on lake catchment dynamics at Lake Mugecuo.

It is becoming increasingly recognised that living organisms can significantly modify geomorphic processes and thus have a strong control on landform dynamics (e.g., Howard and Mitchell, 1985; Naiman et al., 1988; Butler, 1995; Jones et al., 1997a; Crooks, 2002; Cuddington et al., 2007; Gilad et al., 2007). It has been hypothesized that a vegetation cover with a high number of plant species of different growth forms, functional groups, and root characteristics will be more effective at governing interrill erosion processes than a vegetation cover with few species and a lower number of different growth forms (Körner & Spehn 2002). The results of the regression analysis provides evidence to support this hypothesis as it revealed that causal relationships were found between plant functional diversity and specific lake catchment dynamics proxies only when the most significantly correlated PTF was found in conjunction with several other PFTs (section 5.2). The effects of periodic changes in functional diversity were observed in the cross-correlation analysis which showed numerous (lagged) cross-correlations between the coarser particle size proxies with shrub and tree taxa, whilst changes in the finer sediment size proxies correlated with and were driven by an array of plant functional types including sedges and grasses (section 5.2.3). A functional plant cover provides soil stability and is especially important in alpine ecosystems (Pohl et al., 2009). The way the vegetation is spatially distributed along the slopes is an important factor for decreasing the sediment runoff (Lavee et al., 1998 and Calvo-Cases et al., 2003). The spatial and temporal variability in the density, type, and composition of vegetation on banks (Montgomery, 1997 and McBride et al., 2008) or the adjacent floodplain, the channel scale and its location within a watershed (Abernathy and Rutherfurd, 1998, Anderson et al., 2004 and Eaton and Giles, 2009), bank materials (Thorne, 1990 and Dunaway et al., 1994) and the periodicity of large flood events (Nadler and Schumm, 1981 and Friedman et al., 1996) can all influence whether, how, and how much vegetation influences channel form and stability (Labbe et al., 2011). These variabilities likely explain why certain species/taxa correlate with specific
fraction sizes of sediment that are transported into the lake over time. However, it is important to highlight that the results of this study indicate that the plant functional diversity types do not act in isolation in influencing lake catchment dynamics.

The findings of this research indicates that an array of different types of species and/or functional groups play an important role in modulating and landscape evolution across millennial timescales (approximately 1,000-3,000 years). Most importantly these results show that vegetation modulates the lake catchment dynamics data thereby suggesting that vegetation essentially filters the signals of these proxies. Evidence for this scenario was also clear in the use of an independent proxy for catchment-wide erosion rates ($^{10}$Be) which displayed a significant change in erosion rates in the MG1 record at approximately 3.8 kcal.yr BP that was not detected by the traditional proxies for erosion in palaeoenvironmental reconstructions (e.g. Hakanson et al., 1983; Leeder et al., 1992; Zhang Shi et al., 1997; Sperazza et al., 2004; Liu et al., 2008; An et al., 2011). The palaeo-erosion rate samples taken from the mouth of the river delta provided evidence that increased levels of sand in the particle size data in MPZ-6 (Figure 6.1) did not necessarily indicate increased erosion in the catchment. Findings such as this highlight the benefit of adopting a multi-proxy approach into the investigation of catchment dynamics and it is argued here that these proxies are in fact not useful indicators for erosion and deposition in palaeoenvironmental reconstructions. This is an extremely important conclusion of this study and further work is required to clarify its validity.

It was discussed in Chapter 1 that the value of any multi-proxy study clearly rests on the reliability of the proxies used to reconstruct the past environmental conditions; different proxies reflect different environmental factors at a range of spatial scales and consequently show different strengths and weaknesses (Birks and Birks, 2006). The point was made that the combination of such proxies may help to register leads and lags in the reaction of different biotic and abiotic systems to climatic change and thus enhance our understanding of landscape dynamics under a changing climate, including equilibrium or disequilibrium between landscape and climate (Ammann, 1989a). The Lake Mugecuo MG1 record has proven that cosmogenic $^{10}$Be measurements are a useful proxy in palaeoenvironmental reconstruction studies as a method that can provide robust a robust record of denudation changes in a catchment that traditional proxies do not provide (e.g. particle size). There is still a limited understanding of the mechanisms by which biogeomorphic processes contribute to defining earth surface landscape dynamics through reciprocal adjustments between landform dynamics, population dynamics, ecological successions and organism evolution (Dietrich & Perron 2006; Amundson et al 2007, Corenblit and Steiger, 2009; Richter & Mobley, 2009; Reinhardt et al., 2010). This study has provided strong evidence that combining $^{10}$Be measurements of erosion rates with palaeo-vegetation and palaeoenvironmental proxies has the potential to investigate these relationships which may help to predict future landscape change under changing climatic conditions.
6.1.1 Summary of overall findings

Based on the evidence contained in the Lake Mugecuo record and comparison with other regional proxy records, the following conclusions can be drawn;

- The MG1 record provides a high-resolution record of long-term changes in plant functional diversity in this biodiversity hotspot which may have implications for future ecosystem management and it may also inform us about potential future ecological changes under changing climatic conditions in this region.
- This research adopts a unique approach of utilising $^{10}$Be erosion rate measurements in conjunction with ‘classic’ palaeoenvironmental proxies to reconstruct past variation in sediment dynamics in the catchment.
- The MG1 record provides a high-resolution, well-dated recorded of environmental, climatic and vegetational changes in the Lake Mugecuo catchment, Southwest China from the Late-Glacial (12.6 kcal.yr BP) to the Late Holocene (0.0 kcal.yr BP) period.
- Due to the nature of this study and the timescale this study aimed to address, the plant functional type classification scheme utilised in this study provided an appropriate, tested, globally applicable method of assigning fossil pollen taxa to appropriate plant functional groupings which were then used in turn as a method of assessing biomization changes in the Mugecuo catchment, thus inferring broad-scale Holocene EASM variability.
- Long-term changes in climate inferred from the MG1 palaeo-vegetation record broadly follow the changes indicated in the D4 $\delta^{18}$O speleothem reference record of changes in effective moisture patterns in the Holocene for Southwest China.
- The differences in $^{10}$Be concentrations in modern bedrock and fluvial sediments provide evidence of the spatial variations of erosion rates in the catchment and these differences in exposure times indicate the indicate the pathway of sediment through the catchment.
- Environmental shifts captured in the Lake Mugecuo record are primarily driven/mediated by changes in plant functional diversity.
- Vegetation likely mediates the effects of climate change on lake catchment dynamics at Lake Mugecuo based on qualitative analysis of the data.
- An array of different types of species and/or functional groups play an important role in modulating lake catchment dynamics on millennial timescales.
- Vegetation has a filtering effect on the proxies which are traditionally used as indicators for changes in erosion and deposition in palaeoenvironmental reconstructions.
- $^{10}$Be cosmogenic measurements provided supplementary data on past changes in landscape dynamics in the catchment which conventional proxies used in palaeoenvironmental studies were unable to achieve.
6.2 Combining proxies such as $^{10}$Be analysis with palaeoenvironmental proxies- a useful/viable concept?

The limitations of each individual proxy utilised in this study have been demonstrated, but it is these limitations which highlight the importance of a multi-proxy approach to palaeoenvironmental reconstruction studies. Each proxy reflects the environment at its own spatial scale, taking its place in the network of interactions that comprise an ecosystem, thus providing insights into different facets of an ecosystem (Birks & Birks 2006). This study encountered a number of potential problems associated with a) the extraction of sufficient quantities of sand for cosmogenic $^{10}$Be analysis from a lake sediment core and b) obtaining a robust chronology for the sand samples. Due to time constraints and the remote nature of the field site, it was not possible to obtain deeper lake sediment cores for the $^{10}$Be cosmogenic analysis which may have provided a longer-term record of erosion rates for the catchment. However, this study has shown that $^{10}$Be cosmogenic measurements provide supplementary data which conventional proxies used in palaeoenvironmental studies are unable to achieve. It was shown that vegetation actually has a filtering effect on the palaeolimnological proxies utilized in this study.

In terms of the difficulties encountered when attempting to date the sandy material in the shallow water cores, the use of bulk sediment $^{14}$C dating enabled a robust chronology to be constructed for an age-depth profile for the measured palaeo-erosion rates. The application of cosmogenic erosion rates in a depositional lacustrine setting is novel, but this study has shown that it can be achieved in a setting such as this and that it is useful to combine this proxy with other palaeolimnological proxies in order to reconstruct catchment dynamics through time.

6.3 Wider story- the applicability of this research

The Holocene climate and vegetation history of this area in central Sichuan is still poorly investigated and the number of high-resolution records is limited. This study carried out at Lake Mugecuo is one of the first high resolution Holocene palaeoenvironmental and palaeoecological records available for this region in the Hengduan Mountains. It provides a unique insight into the environmental and vegetation changes which have taken place since the Late-Glacial period in this remote part of Southwest China. Our knowledge about the history of past climate change and past ecosystem development and lake ontogeny in Southwest China is steadily increasing (e.g. Jarvis, 1993; Shi et al., 1993; Yu et al 2000; Shen et al., 2005; Herzschuh, 2006; Cook et al., 2012; Jones et al., 2012). However, for such a floristically diverse and biodiversity rich region, robust, high-resolution data is severely lacking in the Hengduan Mountains, a region which is becoming increasingly populated with a sudden surge in tourism in the last 10 years. Despite the recent efforts to study mountain plant diversity in China, there are still significant gaps. This study has provided a long-term record of palaeo-vegetation in response to long-term changes in climate during the Holocene and has shown that using a PFT classification of palaeo-data has the potential to provide a useful means of predicting the assemblage of species’
responses within an ecosystem to environmental conditions or disturbance. The use of a widely accepted classification scheme (Ni et al., 2010; Harrison et al., 2010) removes the need to study every taxon in a pollen record individually and allows for a classification scheme which is globally consistent. The results of this study can thereby be directly comparable to other similar study sites worldwide as it provides a transferable means of adopting a similar multi-proxy approach in a meaningful way.

Ultimately, the relationships between vegetation distribution and climate play an indispensable role in understanding the vulnerability and adaptability of the ecosystems during periods of great climatic change (Bugmann and Pfister 2000; IPCC 2001, 2007; Yu et al. 2006; Ni 2011). By adopting a multi-proxy approach to studying ecosystems similar to the techniques used here, further understanding of the relationships between vegetation, climate and landscape dynamics in alpine regions can potentially be achieved in order to predict how future change will impact them.

6.4 Specific conclusions from this research

In Chapter 1 a set of research questions were proposed as follows:

1. How has a dynamic high alpine lake catchment system in Sichuan Province, Southwestern China responded to EASM variability in terms of key environmental and vegetational shifts since the Early Holocene?
2. Does plant functional diversity mediate the effects of climate on lake catchment dynamics at Lake Mugecuo?
3. Is combining a proxy such as $^{10}$Be analysis with palaeoenvironmental proxies a viable/useful technique/concept in palaeoenvironmental reconstructions?

By addressing these questions during the course of this study, the following conclusions can now be reached:

- The Lake Mugecuo record provides a high resolution palaeoclimatic reconstruction based on natural archives in a region where palaeo-record coverage is presently extremely sparse. It provides a unique insight into the environmental and vegetation changes which took place during the Late-Glacial/Early Holocene transition, and captures the changes which took place throughout the Holocene.
- The Lake Mugecuo record broadly supports palaeoclimatic shifts inferred from palaeo-records across continental Southeast Asia. Pollen studies from this region essentially show a similar vegetation history but the boundary between these changes in often different (Jarvis, 1993; Van Campo and Gasse, 1993; Shen et al., 2005; 2006; Zhu et al., 2009) i.e. there are slight lags in the records likely due to differences in competing climatic influences across sub-regions (Cook et al., 2012).
This study gives a broad-scale picture of the influence of plant functional diversity on lake catchment dynamics through time at Lake Mugecuo, a sensitive high-alpine lake. The results have shown that proxies for catchment dynamics are influenced, albeit weakly, by functional diversity patterns on long-term timescales.

The main advantage of using such a PFT classification scheme is that it is globally applicable and allows for a relatively straightforward comparison of biome reconstructions among regions and continents (Harrison et al., 2010; Ni et al., 2010).

Although we gain a broad-scale picture of the major changes in dominant PFT groupings over time, we may be missing out on the smaller-scale changes in the true plant community dynamics of the catchment.

The apparent lack of information that can be garnered from the MG1 record alone on the potential erosion events that may or may not have occurred in the catchment based on the use of ‘conventional’ proxies through time highlights the potential benefits of combining a palaeolimnological study with the calculation of in-situ erosion rates from $^{10}$Be concentrations— the $^{10}$Be results essentially imply that sediment dynamics proxies such as particle size say very little about erosion.

### 6.5 Suggestions for further work

As there is currently no agreed global PFT scheme an enhanced PFT scheme still remains to be determined. There is a need for a bio-climatically oriented global PFT classification which is complemented by information on traits related to dynamic processes within ecosystems (competition for light, water and nutrients, successional dynamics and the disturbance regime) (Harrison et al., 2010). A new PFT scheme should further exploit the availability of quantitative trait information to go beyond the classical dichotomies (evergreen/deciduous, sclerophyll/malacophyll, etc.) and recognize the existence of continuous trait variation in measurable quantities, and the ubiquity of trade-offs among such quantities (Wright et al. 2004). Further study is required to address the potential trade-offs between looking at small and larger-scale dynamics.

The application of remote sensing in ecological research provides a novel method of assessing vegetation based on physiognomic and structural characters rather than floristic composition (Armitage et al., 2000). PFTs represent an extension of this characterisation of vegetation and the combination of these two techniques have the potential to assess plant community dynamics on a narrower scale than is currently possible (see Duckworth et al., 2001). Interpreting palynological records from the late Pleistocene and Holocene requires an understanding of the relationship between modern vegetation and the pollen rain (Pisaric et al., 2000). In a future study, remote
sensing could be used to determine the modern-day spread of vegetation whilst employing transfer functions on palaeo-vegetation data in order to assess how stands of vegetation have changed through time and thus how the ‘spread’ of plant functional types and plant community dynamics potentially change through time.

- A similar study by Grischott et al (2013) dated sediment cores from a fluvial fan in an alpine lake in the Alps using cosmogenic dating which ranged back to 2320 years. This new data in conjunction with the data obtained at Lake Mugecuo indicates that in order to get longer-term palaeo-erosion rates from lacustrine settings, more knowledge needs to be obtained on the history of a catchment in order to assess how a river may have meandered back and forth over time and thus where the older sand deposits may lie in the landscape. Techniques such as echo depth sounding and shallow seismic sub-bottom profiling (e.g. Dietze et al., 2010) could potentially provide this information in order to provide a more robust method for the selection of appropriate coring sites in order to obtain the material required for longer-term $^{10}$Be measurements.
Appendix A.1: Additional information for Chapter 2 (Methods)

A.1.1 The Bayesian P_sequence depositional model used in Oxcal v.4.17 (MG1 is used as an example)

P_Sequence("",1000,50) {
    Boundary("bottom") { };
    R_Date("381cm", 10650,40) { z=381; };
    R_Date("343cm", 9392,42) { z=343; };
    R_Date("341cm", 10319,40) { z=341; };
    R_Date("338cm", 9120,40){ z=338;};
    R_Date("334cm", 8128,59) { z=334; };
    R_Date("322cm", 10271,39) { z=322; };
    R_Date("290cm", 8840,120) {  z=290; };
    R_Date("248cm", 8000,40) {  z=248; };
    R_Date("200cm", 6180,40) {  z=200; };
    R_Date("150cm", 4180,30) {  z=150; };
    R_Date("100cm", 2370,30) { z=100; };
    R_Date("50cm",  1780,30) { z=50; };
    Boundary("top") { z=1; };
};

A.1.2 Cosmogenic \(^{10}\)Be Sample Descriptions

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Site Description</th>
<th>Co-ordinates (DD) &amp; Altitude (m)</th>
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<tbody>
<tr>
<td>MGCR</td>
<td>Multiple bedrock samples (n=10) were collected from a single hill-slope exposure on the north eastern side of the lake catchment</td>
<td>30.155, 101.55 3,831</td>
</tr>
<tr>
<td>MGDR</td>
<td>Modern river sediment (0.25-0.5mm) was collected from the surface layer of the small delta below the river outlet which is the major source of water to the lake.</td>
<td>30.159, 101.54 3,788</td>
</tr>
<tr>
<td>MGMR</td>
<td>Modern river sediment (0.25-0.5mm) was collected from the surface layer of the small delta below the second greatest source of water to the lake.</td>
<td>30.155, 101.53 3,790</td>
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</tbody>
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A.1.3 Calculations of the scaling factors for $^{10}$Be erosion-rate measurements

The following script calculates the geographic scaling factor for cosmogenic-nuclide production as a function of site latitude and atmospheric pressure, according to Stone (2000) and Balco (2001).

% input data
dem = xlsread('chinadem.xlsx');
dem(dem==0)=NaN;
dem(dem==-9999)=NaN;
[m,n]=size(dem);

% Syntax: pressure = stdatm(elevation)
% Units: elevation in m; pressure in hPa; accepts vector arguments
% This function converts elevation to atmospheric pressure according
% to the "standard atmosphere" (cf. CRC Handbook of Chem and Phys).
% function out = stone2000(lat,P,Fsp); Calculates the geographic scaling factor for cosmogenic-nuclide
% production as
% a function of site latitude and atmospheric pressure, according to:
% Stone, J., 2000, Air Pressure and Cosmogenic Isotope Production. JGR 105:B10,
% p. 23753.
% Syntax: scalingfactor = stone2000(latitude,pressure,fsp)
% Units:
% latitude in decimal degrees
% pressure in hPa
% Fsp is the fraction (between 0 and 1) of production at sea level
% and high latitude due to spallation (as opposed to muons).
% This argument is optional and defaults to 0.978, which is the value
% used by Stone (2000) for Be-10. The corresponding value for Al-26
% is 0.974. Note that using 0.844 for Be-10 and 0.826 for Al-26 will
% closely reproduce the Lal, 1991 scaling factors as long as the standard
% atmosphere is used to convert sample elevation to atmospheric pressure.
% Also note that this function will yield the scaling factor for spallation
% only when Fsp=1; and that for muons only when Fsp=0.
% Elevation can be converted to pressure with the functions
% stdatm.m (general use) and antatm.m (Antarctica).
% Vector arguments are OK. All arguments must be the same size.
% Written by Greg Balco -- UW Cosmogenic Nuclide Lab
% balcs@u.washington.edu
% First version, Feb 2001
% Checked March, 2006
% Part of the CRONUS-Earth online calculators:
% http://hess.ess.washington.edu/math
% Copyright 2001-2007, University of Washington
% All rights reserved
% Developed in part with funding from the National Science Foundation.
% This program is free software; you can redistribute it and/or modify
% it under the terms of the GNU General Public License, version 2,
% as published by the Free Software Foundation (www.fsf.org).

% define variables
gmr = -0.03417;
Ts = 288.15;
dtdz = 0.0065;
Ps = 1013.25;
shielding=0.99;

out_1 = Ps.*exp((gmr/dtdz).*(log(Ts) - log(Ts - (dem.*dtdz))));
if size(out_1,1)>size(out_1,2);
    out = out_1';
else
    out = out_1;
end;
P = out_1;
lat = ones(length(P),1).*30.15;
long = 101.84;
Fsp = 0.97;

function out = stone2000(lat,P, Fsp);

%% check for obvious errors
if ~isempty(find(abs(lat) > 90));
    error('Latitudes below 90, please');
end;

if length(lat) ~= length(P);
    error('Vectors the same size, please');
end;

if nargin == 2;
    Fsp = 0.978;
end;

%% spallation scaling factor
lat = abs(lat);
lat(find(lat > 60)) = (zeros(size(find(lat > 60))) + 60); %makes all lats above 60 = 60

a = [31.8518 34.3699 40.3153 42.0983 56.7733 69.0720 71.8733];
b = [250.3193 258.4759 308.9894 512.6857 649.1343 832.4566 863.1927];
c = [-0.083393 -0.089807 -0.106248 -0.120551 -0.160859 -0.199252 -0.207069];
d = [7.4260e-5 7.9457e-5 9.4508e-5 1.1752e-4 1.5463e-4 1.9391e-4 2.0127e-4];
e = [-2.2397e-8 -2.3697e-8 -2.8234e-8 -3.8809e-8 -5.0330e-8 -6.3653e-8 -6.6043e-8];
ilats = [0 10 20 30 40 50 60];

dd = 1;
while dd <= m;
a1 = interp1(ilats,a, lat(dd));
b1 = interp1(ilats,b, lat(dd));
c1 = interp1(ilats,c, lat(dd));
d1 = interp1(ilats,d, lat(dd));
e1 = interp1(ilats,e, lat(dd));
S(dd,1:n) = a1 + (b1 .* exp(P(dd,:)./(-150))) + (c1.*P(dd,:)) + (d1.*(P(dd,:).^2)) + (e1.*(P(dd,:).^3));
end;

%% muon scaling factor
mk = [0.587 0.600 0.678 0.833 0.933 1.000 1.000];

dd = 1;
while dd <= m;
    M = interp1(ilats,mk, lat(dd));
    muon(dd,1:n) = M .* exp((1013.25 - P(dd,:))./242);

dd = dd+1;
end;

%% combined scaling factor

scalingfactor = ((S .*Fsp) + (muon .*((1 - Fsp)))* shielding;

Po=scalingfactor.*4.5;
meanPo=nanmean(Po(:))

n=60/((0.5*2.65)/160+log(2)/1387000);
m=(10^-13*0.000200*(6.022*10^23)/10)/n
A.1.4 CANOCO v.4.5.2 DCA/PCA metadata

The following metadata was generated during the statistical analysis of the Lake Mugecuo pollen dataset including taxa with abundances of >0.5% using CANOCO v.4.5.2 (ter Braak and Smilauer, 2003).

Stage 1: DCA
No samples omitted
Number of samples  191
Number of species  15
Number of occurrences  2728
Squareroot-transformation of species data
No species-weights specified
No sample-weights specified
donweighting of rare species
Final species weights applied (weight*downweight)

<table>
<thead>
<tr>
<th>1.000</th>
<th>1.000</th>
<th>1.000</th>
<th>1.000</th>
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<th>1.000</th>
<th>1.000</th>
<th>1.000</th>
<th>1.000</th>
<th>1.000</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Axes are rescaled
No. of active samples: 191
No. of passive samples: 0
No. of active species: 15

Total inertia in species data =
Sum of all eigenvalues of CA = 0.12560

*** Summary ****

<table>
<thead>
<tr>
<th>Axes</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>Total inertia</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0.053</td>
<td>0.015</td>
<td>0.009</td>
<td>0.007</td>
<td>0.126</td>
</tr>
<tr>
<td></td>
<td>0.745</td>
<td>0.713</td>
<td>0.466</td>
<td>0.466</td>
<td></td>
</tr>
<tr>
<td>Cumulative percentage variance of species data</td>
<td>41.9</td>
<td>54.2</td>
<td>61.7</td>
<td>67.3</td>
<td></td>
</tr>
<tr>
<td>Sum of all eigenvalues</td>
<td>0.126</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Stage 2: PCA
No samples omitted
Number of samples  191
Number of species  15
Number of occurrences  2728
Squareroot-transformation of species data
No species-weights specified
No sample-weights specified
Centering/standardization by species = 3
Centering/standardization by samples = 0

No. of active samples: 191
No. of passive samples: 0
No. of active species: 15

Total sum of squares in species data = 2865.00
Total standard deviation in species data TAU = 1.00000
(after species centering/standardization, if in force)

**** Summary ****

<table>
<thead>
<tr>
<th>Axes</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>Total variance</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0.315</td>
<td>0.178</td>
<td>0.089</td>
<td>0.081</td>
<td>1.000</td>
</tr>
<tr>
<td>Cumulative percentage variance of species data</td>
<td>31.5</td>
<td>49.3</td>
<td>58.1</td>
<td>66.3</td>
<td></td>
</tr>
<tr>
<td>Sum of all eigenvalues</td>
<td>1.000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
A.1.5 The D4 δ²¹⁸O record

The D4 δ²¹⁸O record was interpolated to an evenly spaced data set with a resolution of 1 year and was then transformed using a very low degree of smoothing using the filter function in using the following script to give a moving average of approx. 66 years (to match the resolution of the pollen record every 2cm= ~66 years).

Dongge = xlsread ('Dongge.xls'); %Need to make sure there are no duplicates of data
a = Dongge(:,1);
b = Dongge(:,2);
xi = 0:1:9470; %new time axis which runs from 0-12500kyr with 1yr intervals
yi = interp1(a,b,xi); %interpolation formula
plot(a,b,'k',xi,yi,'c')

%Filter Function
%data = yi'; %transposes the data back to a column
yi = yi';
plot(a,b,'k',xi,yi,'c')

windowSize = 60; %the size of moving window 60 here because pollen record resolution
smoothdataD=filter(ones(1,windowSize)/windowSize,1,yi);
plot(a,b,'k'); hold on; plot(xi,smoothdataD,'c');

% NB need to sort out beginning of record somehow (can’t just add to
% beginning of record as values need to be distinct
% Downsampling resamples input @ lower rate by deleting samples
% Decrease the sampling rate of x by keeping every n-th sample starting
% with the first sample. x can be a vector or matrix, if x is a matrix each
% column is considered a separate sequence
downsamplingD = downsample (smoothdataD,60);
Appendix A.2: Additional information for Chapter 5 (Discussion)

A.2.1: Spearman Rank Correlation Coefficients

Matlab Script:

```matlab
data = xlsread('StdevAllData.xlsx');
x = data(:,2);
y = data(:,3);
rho = corr(x, y, 'type', 'Spearman');
```

Supplementary Results:

Spearman's correlation coefficients (R²) for each of the five particle size categories and the Dongge Cave speleothem data. Statistically significant correlations are highlighted in bold italics. All data were smoothed, interpolated and down-sampled to create a complete dataset record at 66 year sample intervals. These data were de-trended, centred and expressed in terms of standard deviation units.

<table>
<thead>
<tr>
<th>Proxy</th>
<th>D4 5180</th>
</tr>
</thead>
<tbody>
<tr>
<td>PS &lt; 4.00 μm</td>
<td>0.1477</td>
</tr>
<tr>
<td>PS &gt;4μm &lt; 16μm</td>
<td>0.2413</td>
</tr>
<tr>
<td>PS &gt;16μm &lt;32 μm</td>
<td>-0.3937</td>
</tr>
<tr>
<td>PS &gt;32μm &lt;64μm</td>
<td>-0.15</td>
</tr>
<tr>
<td>PS &gt;64μm</td>
<td>0.01</td>
</tr>
</tbody>
</table>
A.2.2 Cross-Correlation Analysis

Xcorr Matlab Script:

```matlab
data = xlsread ('StDevAllData.xlsx');
x = data(:,5);
y = data(:,15);
%lags = ((length(x)-1)^2-1:length(x)-1);
precision = 66;
nloops=5000;
n=length(x);%number of lags = n
corrcolumn=ones(1,n*2-1).*NaN;%create empty column to fill with corr data
datamatrix=ones(nloops,n*2-1).*NaN;
for i= 1:nloops
    % randomise data
    r1=randperm(n);%generate random numbers between 1 and no. rows
    r2=randperm(n);%generate random numbers between 1 and no. rows
    xr=sortrows([x,r1],2);%randomise data column-use sort to rearrange row indexing
    yr=sortrows([y,r2],2);%randomise data column-use sort to rearrange row indexing
    % estimate correlation for n lags
    datamatrix(i,:)=xcorr(xr(:,1),yr(:,1),'coeff');%base correlation on the first n cells to keep sample size constant
end
max95=std(datamatrix)*2;%calc 2-sigma value
min05=std(datamatrix)*-2;
% we hypothesise that clamate is the forcing thus climate must be the first
% term in xcorr expression
plot(lags*precision,xcorr(x,y,'coeff'));
hold on
plot(lags*precision,max95,'r');
hold on
plot(lags*precision,min05,'r');
% axis([min(lags*-4) max(lags*4) -0.5 0.5])
% title('xcorrelation between warm season precipitation and sediment source area:pca1')
title('xcorrelation between Climate and PFTs')
xlabel('Lagged time (yrs)');ylabel('x-correlation');
finalresult=[lags*precision;xcorr(x,y,'coeff');max95;min05];
%plot(lags*precision,xcorr(x,y,'coeff'));
```

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Supplementary Results:

Cross-correlation showing the lagged response of the particle-size fractions < 4.00 μm, >4μm < 16μm, >16μm <32 μm, >32μm <64μm and >64μm to changes in the regional climate record inferred from the D4 δ18O at Dongge Cave. Weak but significant lagged landscape responses are observed on millennial timescales. Significance (in red) is denoted by 95% confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-correlation showing the lagged response of the nine plant functional type classes to changes in the regional climate record inferred from the smoothed D4 δ18O at Dongge Cave. Weak but significant lagged landscape responses are observed on millennial timescales. Significance (in red) is denoted by 95% confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-correlation showing the lagged response of PS class <4μm to changes in the plant functional diversity record. Weak but significant lagged landscape responses are observed on centennial timescales. Significance (in red) is denoted by 95% confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-correlation showing the lagged response of PS class >4µm < 16µm to changes in plant functional diversity. Weak but significant lagged landscape responses are observed on millennial timescales at. Significance (in red) is denoted by 95% confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-correlation showing the lagged response of PS class >16\(\mu m\) <32 \(\mu m\) to changes in plant functional diversity. Weak but significant lagged landscape responses are observed on millennial timescales. Significance (in red) is denoted by 95% confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-correlation showing the lagged response of PS class >32µm <64µm to changes in plant functional diversity. Weak but significant lagged landscape responses are observed on centennial (PFT2) and millennial timescales. Significance (in red) is denoted by 95% confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-correlation showing the lagged response of PS class >64µm to changes in plant functional diversity. Weak but significant lagged landscape responses are observed on centennial (PFT2, PFT7) and millennial timescales (1254 years). Significance (in red) is denoted by 95% confidence lines beyond which correlations are unlikely to arise purely by chance.
A.2.3 Lomb-Scargle Scargle Fourier transform

Lomb Matlab Script:

```matlab
function [f,P,prob] = lomb(t,h,ofac,hifac)
% LOMB(T,H,OFAC,HIFAC) computes the Lomb normalized periodogram (spectral
% power as a function of frequency) of a sequence of N data points H,
% sampled at times T, which are not necessarily evenly spaced. T and H must
% be vectors of equal size. The routine will calculate the spectral power
% for an increasing sequence of frequencies (in reciprocal units of the
% time array T) up to HIFAC times the average Nyquist frequency, with an
% oversampling factor of OFAC (typically >= 4).
% %
% % The returned values are arrays of frequencies considered (f), the
% % associated spectral power (P) and estimated significance of the power
% % values (prob). Note: the significance returned is the false alarm
% % probability of the null hypothesis, i.e. that the data is composed of
% % independent gaussian random variables. Low probability values indicate a
% % high degree of significance in the associated periodic signal.
% %
% % Although this implementation is based on that described in Press,
% % Teukolsky, et al. Numerical Recipes in C, section 13.8, rather than using
% % trigonometric recurrences, this takes advantage of MATLAB's array
% % operators to calculate the exact spectral power as defined in equation
% % 13.8.4 on page 577. This may cause memory issues for large data sets and
% % frequency ranges.

%data = xlsread ('StDevAllData.xlsx');
h = data (:,16);
t = data (:,1);
%sample length and time span
N = length(h);
T = max(t) - min(t);
ofac = 4; hifac = 1;
%mean and variance
mu = mean(h);
s2 = var(h);
%calculate sampling frequencies
f = (1/(T*ofac):1/(T*ofac):hifac*N/(2*T)).';

%angular frequencies and constant offsets
w = 2*pi';
tau = atan2(sum(sin(2*w*t.'),2),sum(cos(2*w*t.'),2))./(2*w);

%spectral power
cterm = cos(w*t.' - repmat(w.*tau,1,length(t)));
sterm = sin(w*t.' - repmat(w.*tau,1,length(t)));
P = (sum(cterm*diag(h-mu),2).^2./sum(cterm.^2,2) + ...
   sum(sterm*diag(h-mu),2).^2./sum(sterm.^2,2))/(2*s2);

%estimate of the number of independent frequencies
M=2*length(f)/ofac;

%statistical significance of power
prob = M*exp(-P);
inds = prob > 0.01;
prob(inds) = 1-(1-exp(-P(inds))).^M;

plotyy(f,P,f,prob)
xlabel('Frequency')
ylabel('Power')
title('Lomb-Scargle Power Spectrum')
```

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A.2.4 Cross-Spectral Analysis

Matlab Script:

(Welch’s Method)
%%% Variables
clear f power_frequency max95 pxy variable1 nfft freq datamatrix random1 random2 rand_variable1
PCA_casty_xspectra_results_for_excel
data = xlsread ('StDevAllData.xlsx');
x = data (:,:2);
y = data (:,:3);
variable1=x;% single column of data
t_variable2=y;% define this variable at beginning of main loop below
freq=1/66;
nfft=256;
nloops=5000;
sig_level=95;

%%% ********
% for ii=2:13
% variable2=casty_300yr_precip4_stddev(:,ii);% multiple columns of data

%%% calculate significance levels
% datamatrix =[];
% for ii=1:nloops
% random1=abs(detrend(rand(n,1)));
% random2=abs(detrend(rand(n,1)));
% rand_variable2=sortrows([variable2,random1],2);%randomise data column
% rand_variable1=sortrows([variable1,random2],2);%randomise data column
% [pxy,~]=cpsd(rand_variable2(:,1),rand_variable1(:,1),[],0,nfft,freq);
% lgth=length(pxy);
% datamatrix(i,1:lgth)=abs(pxy)';
% end
% max95=std(datamatrix)*2;%calc 2-sigma sig. value

max95= prctile(datamatrix,sig_level)';%randomised data are not normally distributed
max5= prctile(datamatrix,100-sig_level)';
[pxy, f]=cpsd(variable2,variable1,[],0,nfft,freq);
power_frequency=abs(pxy);
plot(f,power_frequency),grid
hold on
plot(f,max95,'r')%sig line
plot(f,max5,'r')%sig line
title('xxxxxx')
Supplementary Results:

Cross-spectra between the particle size fraction < 4.00 μm and the plant functional types that proved non-significant. All data was detrended, centred and expressed in terms of standard deviation units. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.

Cross-spectra between the particle size fractions and the D4 δ180 record at Dongge Cave that proved non-significant. All data was detrended, centred and expressed in terms of standard deviation units. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-spectra between the particle size fraction < 4.00 μm and the plant functional types that proved non-significant. All data was detrended, centred and expressed in terms of standard deviation units. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-spectra between the particle size fraction >4µm < 16µm and plant functional diversity that proved non-significant. All data was detrended, centred and expressed in terms of standard deviation units. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-spectra between the particle size fraction >16µm <32 µm and the plant functional types that proved non-significant. All data was detrended, centred and expressed in terms of standard deviation units. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-spectra between the particle size fraction >32µm <64µm and plant functional diversity that proved non-significant. All data was detrended, centred and expressed in terms of standard deviation units. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
Cross-spectra between the particle size fraction >64µm and plant functional diversity that proved non-significant. All data was detrended, centred and expressed in terms of standard deviation units. For PFT2 a lag of 198 years was taken away, for PFT7 a lag of 264 years was taken off the data. Significance (in red) is denoted by 95% (upper red line) and 5% (lower red line) confidence lines beyond which correlations are unlikely to arise purely by chance.
A.2.5 Stepwise Regression Analysis:

Stepwise Regression Matlab Script:

data = xlsread('StDevAllData.xlsx'); % Need to make sure there are no duplicates of data
da1 = data(:,17); % 2 cm resolution based on prelim age-depth chronology
da2 = data(:,18);
da3 = data(:,19);
da4 = data(:,20);
da5 = data(:,21);
da6 = data(:,22);
da7 = data(:,23);
da8 = data(:,24);
da9 = data(:,25);
da10 = data(:,26);
da11 = data(:,27);
da12 = data(:,28);
da13 = data(:,29);
da14 = data(:,30);
da15 = data(:,31);
y = data(:,12);
x = [a1,a2,a3,a4,a5,a6,a7,a8,a9,a10,a11,a12,a13,a14,a15]; % Import independent variables and convert them to a matrix
stepwise(x,y);
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