A palaeoenvironmental history of the southern Bekaa Valley and the Lebanon Mountains, Lebanon during the last glacial period (~ 112-35 ka BP)

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School of Geography and the Environment

Thesis Submitted to the University of Oxford
For the Degree of Doctor of Philosophy
Hilary Term 2014

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Abstract

The remnant montane forests of southern Lebanon provide habitat to a rich diversity of endemic, rare and threatened plant and animal species. Prior to the intensification of human activity during the mid-Holocene period these forests covered most of the Lebanon and Anti-Lebanon Mountain ranges. What is not known is how the dynamics of these forests changed in response to major shifts in climate. The aim of this DPhil thesis was to document changes in vegetation composition in southern Lebanon during the last glacial period (~112 to 35 ka BP). To achieve this aim, a palynological study was conducted on a sedimentary sequence recovered from Aammiq Wetland located in the Bekaa Valley, Lebanon. Fossil pollen was analysed to reconstruct variability in vegetation composition and abundance. Energy dispersive spectroscopy, magnetic susceptibility and loss-on-ignition were used to reconstruct the geomorphological and hydrological processes acting within the wetland and its watershed. Macro and microcharcoal were analysed to reconstruct local and regional fire histories respectively. A chronology was established for this sequence using a combination of radiocarbon and optically stimulated luminescence dating (OSL) techniques.

The results indicate significant variability in the composition and abundance of vegetation through the time interval ~ 112 to ~ 35 ka BP closely aligned in time to global and regional climatic events. Changes in the chemical and physical properties of the sediments indicate significant variability in landscape erosion processes and hydrological conditions in the watershed and within the wetland over the similar interval in time. Peaks in local and regional fire activity were concurrent with periods of precession minima and Dansgaard-Oeschger warming events and had a significant impact on the composition of vegetation within the Aammiq region. These findings provide the first account of environmental change in the southern Bekaa Valley and adjacent slopes of the Lebanon Mountains during the last glacial period.

Acknowledgments

There are two people who I wish to thank foremost without their help my D.Phil. would not have come to its conclusion. My mum Valerie Quill who died in the summer before I started my undergraduate degree at Oxford who was as proud as any parent could be on finding out that I had gained a place to read geography at Harris Manchester College, Oxford. Also just for being my mum who did an incredible job of raising two kids on her own whilst doing the most awful jobs to keep food on the table and shoes on our feet. To my wife Lizzy I owe a huge debt of gratitude for firstly putting up with me that is an achievement in itself and secondly for all the help proof reading my draft papers (again and again) and trying to correct my awful grammar, which quite frankly is just as appalling now as when I started at Oxford. To my two children Anna and Ben I am grateful for the sleepless nights, grey hairs and a house full of laughter and smiles.

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Chapter 1: Introduction

> *"There is something fascinating about science. One gets such wholesale returns of conjecture out of such a trifling investment in fact"*
>
> Mark Twain (1874) On Scientific Inquiry

Global change over the last 800,000 years has been characterised by a series of eight glacial-interglacial cycles (Raymo et al., 1998, Augustin et al., 2004). The theory that these cycles were the result of variations in Earth's orbital parameters in relationship to the sun was first proposed by Croll (1867 a, b), and elaborated by Milankovitch (1941). In 1976 the theory was proven through the work of Hays, Imbrie and Shackleton (Hays et al., 1976) who observed orbital cycles in a 500,000 year record of oxygen isotopes from a marine sediment core. They showed that changes in the oxygen isotopic composition of planktonic foraminifera occurred in phase with variations in Earth's orbital parameters in cycles of 100 ka (eccentricity), 42 ka (obliquity) and 23 ka (precession). These cycles relate to variations in the amount of insolation received by the Earth. The dominant climate change occurred in phase with eccentricity (Hays et al., 1976) which is associated with the shift from an interglacial to glacial period (Sigman and Boyle, 2000, Berger, 2001).

The glacial periods were characterised by the development of continental ice sheets in the northern hemisphere at high latitudes (Hays et al., 1976, Ruddiman and McIntyre, 1979, Birchfield et al., 1981). These ice sheets interacted with the rest of the climate system directly and indirectly by influencing sea surface temperatures, ocean circulation, continental water balances, atmospheric circulation and vegetation at the global scale (Clark et al., 1999). In un-glaciated parts of northern Europe the landscape was largely treeless steppe-tundra during the late-Pleistocene glacial periods (Van Andel and Tzedakis, 1996, West, 2000). However, the climate of the last glacial period in northern Europe was not uniformly cold and dry but instead was inter-dispersed by ‘interstadial’ warm periods of millennial and centennial duration (Fletcher et al., 2010), which facilitated the growth of open woodland (Woillard, 1978, Behre, 1989, Müller et al., 2003). In central and eastern Europe, at least some of the landscape was populated by trees even during the coldest stages of the last glacial (Willis et al., 2000, Willis and Van Andel, 2004). In southern Europe, various studies indicate that the vegetation mosaic was primarily comprised of steppic vegetation during cold/dry stadials and that forest expanded during interstadials (Tzedakis, 1994, Allen et al., 1999, Sánchez Goñi et al., 2002, Tzedakis et al., 2004, Margari et al., 2009, Lezine et al., 2010). However, to date, knowledge on the composition of vegetation in southern Europe is patchy due to a smaller number of suitable palaeoecological sites spanning this interval in time and with particular gaps in southern Italy, Turkey and the Levant in the eastern Mediterranean (Fletcher et al., 2010).

This thesis reports new palaeoenvironmental data spanning ~ 112 to ~ 35 ka BP that was recovered from Aammiq Wetland, Lebanon (Fig.1.1) in the eastern Mediterranean. These data are used to reconstruct the environmental processes that occurred within the wetland and its surrounding watershed during the shift between the last interglacial and glacial period and through the last glacial period up to 35 ka BP, with a particular focus on the dynamics on vegetation in response to global and regional climate forcing. Changes in chemical and physical properties of the sediments are used to ascertain how geomorphological and hydrological processes on the landscape were affected by climate variability and how they in turn affected the state of the water body and vegetation in the wetland and surrounding areas. The local and regional fire histories are also reconstructed in order to gain insight into the complex relationship between peak fire events, climate change and vegetation composition.

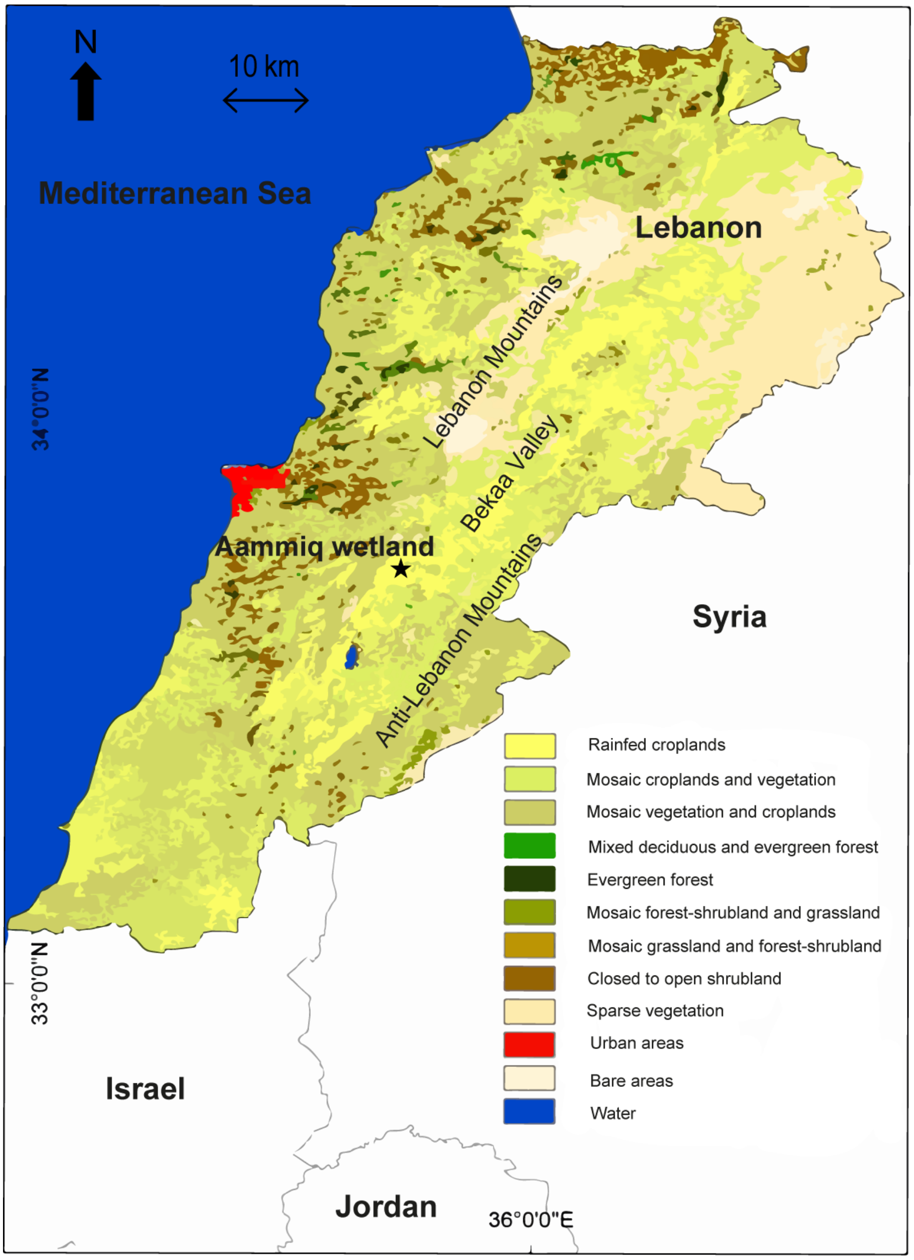

Figure 1.1 Land cover map of Lebanon constructed using information from GLOBCOVER (Bicheron et al., 2009) also showing location of Aammiq Wetland.

1.1 Current knowledge on the history of Lebanon's environment

Human disturbance as a driver of ecological change (mid-Holocene ~ 6 ka BP to present)

The Lebanese landscape has historically been subjected to extensive human activity (Fady et al., 2008, Asmar, 2011). As early as the 3rd Millennium BC, the Phoenicians and Egyptians travelled to Lebanon to extract timber for ship building and construction (Talhouk et al., 2001). The wood of *Cedrus libani* (Cedar of Lebanon) was particularly prized for its durability, density, colour and fragrance (Fady et al., 2008). Its use is mentioned several times in the Old Testament and in one of the oldest known recorded legends, the Epic of Gilgamesh, 2000 BC (Khuri et al., 2000). The cutting of the cedar forests continued under Roman tutelage (Meiggs, 1982) and then under the Ottoman Empire (Talhouk et al., 2001). During British colonial rule in the First World War, cedars were felled for construction of the Aleppo to Baghdad railway (Talhouk et al., 2001) and again during World War II for the Tripoli to Haifa railway (Mikesell, 1969).

It is estimated that at the onset of the Holocene (~ 11.5 ka BP), long before the historical exploitation of these forests, cedar trees covered up to 500,000 ha of Lebanon's upland areas (Alptekin et al., 1997). In modern times the area covered by cedar forests has declined further due to continued extraction, land clearance for agriculture and housing (Talhouk et al., 2001) and to overgrazing by goats and cattle (Beals, 1965) that remove young saplings. A census of the extent of *Cedrus libani's* coverage in the 1980's concluded that it had been reduced to just over 2000 ha (Khouzami, 1994). By 2003, this acreage had been further reduced to 12 scattered remnant patches (Sattout et al., 2007) (see Fig. 1.2), covering just 1135 ha or 0.86% of the total area of forested land in Lebanon (Lichaa El-Khoury and Bakhos, 2003). The decline of this emblematic tree species (Asmar, 2011) is often viewed as a symbol of the environmental degradation of this country.

Anthropogenic degradation of the natural environment in Lebanon has not been restricted to the upland primary forests. During Medieval times the central and southern parts of the Bekaa Valley are known to have been covered by lakes, swamps and seasonally flooded marshes (Scott, 1995). These aquatic/semi-aquatic environments have dried out due

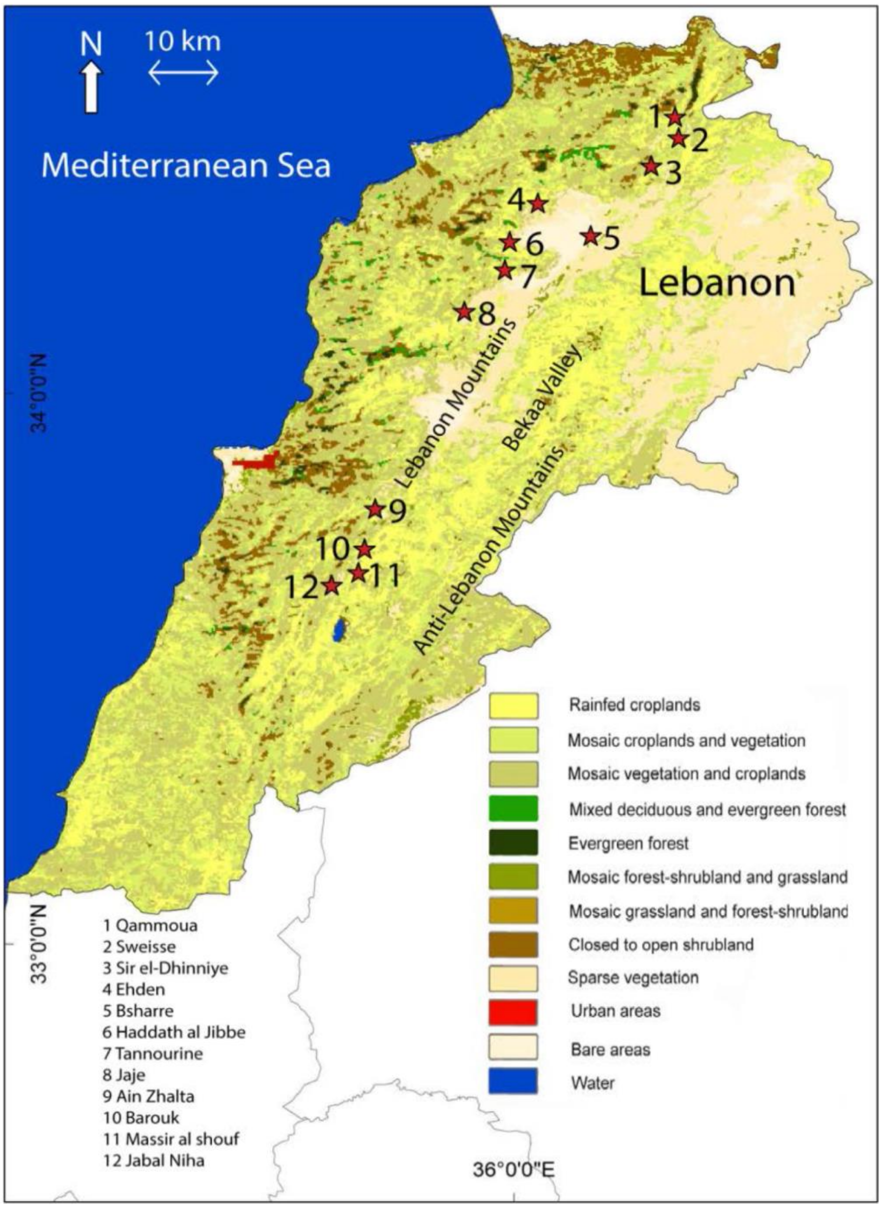

Figure 1.2 Land cover map of Lebanon constructed using information from GLOBCOVER (Bicheron et al., 2009) showing location of remnant populations of *Cedrus libani* (Cedar of Lebanon).

to natural climatic variation and this has been greatly exacerbated by the conversion of land for agriculture. Today Aammiq Wetland is the last remnant of these once extensive wetlands (Walley, 1996) and Lebanon's only surviving wetland of high conservation status (RAMSAR, 2001). Aammiq provides the last refuge for numerous plants and animals that were probably more numerous during previous periods of moister climatic conditions in the Holocene and Pleistocene (Walley, 1996).

Climate as a driver of ecological change during the early to mid-Holocene MIS 1 (~ 11.5 to ~ 6 ka BP) prior to human transformation of the environment

During the early-Holocene (~ 11.5 to ~ 7 kyr cal. BP) there is no evidence of large scale human disturbance of the natural environment in Lebanon (Hajar et al., 2009). However, there were major changes in upland forest and wetland ecosystems (Hajar et al., 2009) and this has been attributed to climatic change that occurred at this time. Previous pollen analysis conducted at Aammiq Wetland (see Fig. 1.3) that covers the late-deglaciation and Holocene (~ 14 kyr cal. BP to present day) suggests that the upland areas of southern/central Lebanon during the early-Holocene were covered by open oak forests (Table 1.1) with small populations of *Cedrus libani* within the forest mosaic (Hajar et al., 2008). At Yammoûneh Basin in northern Lebanon (Fig. 1.4), the presence of deciduous oak forests (Fig. 1.5) in conjunction with the deposition of white lacustrine marl during the early-Holocene indicates that the climate was wet and relatively warm (Gasse et al., 2011). Low $\delta^{18}O$ and $\delta^{13}C$ values of carbonate in spelothems from Jeita Cave, western Lebanon (Fig. 1.4) suggest that this was the wettest period of the Holocene (Verheyden et al., 2008).

High levels of precipitation during the early 'Holocene climatic optimum' were not confined to Lebanon, but persisted across the entire eastern Mediterranean (Robinson et al., 2006). The speleothem record from Soreq Cave, Israel (Fig. 1.4) indicates that the early-Holocene was the wettest period since the onset of deglaciation, with almost twice the amount of annual rainfall as in the present-day (Bar-Matthews et al., 1999, Bar-Matthews et al., 2000). The marine sedimentary record from the Mediterranean Basin reflects similar amelioration of the climate, as indicated by the deposition of a sapropel at 9.3 ± 0.13 to 8.57

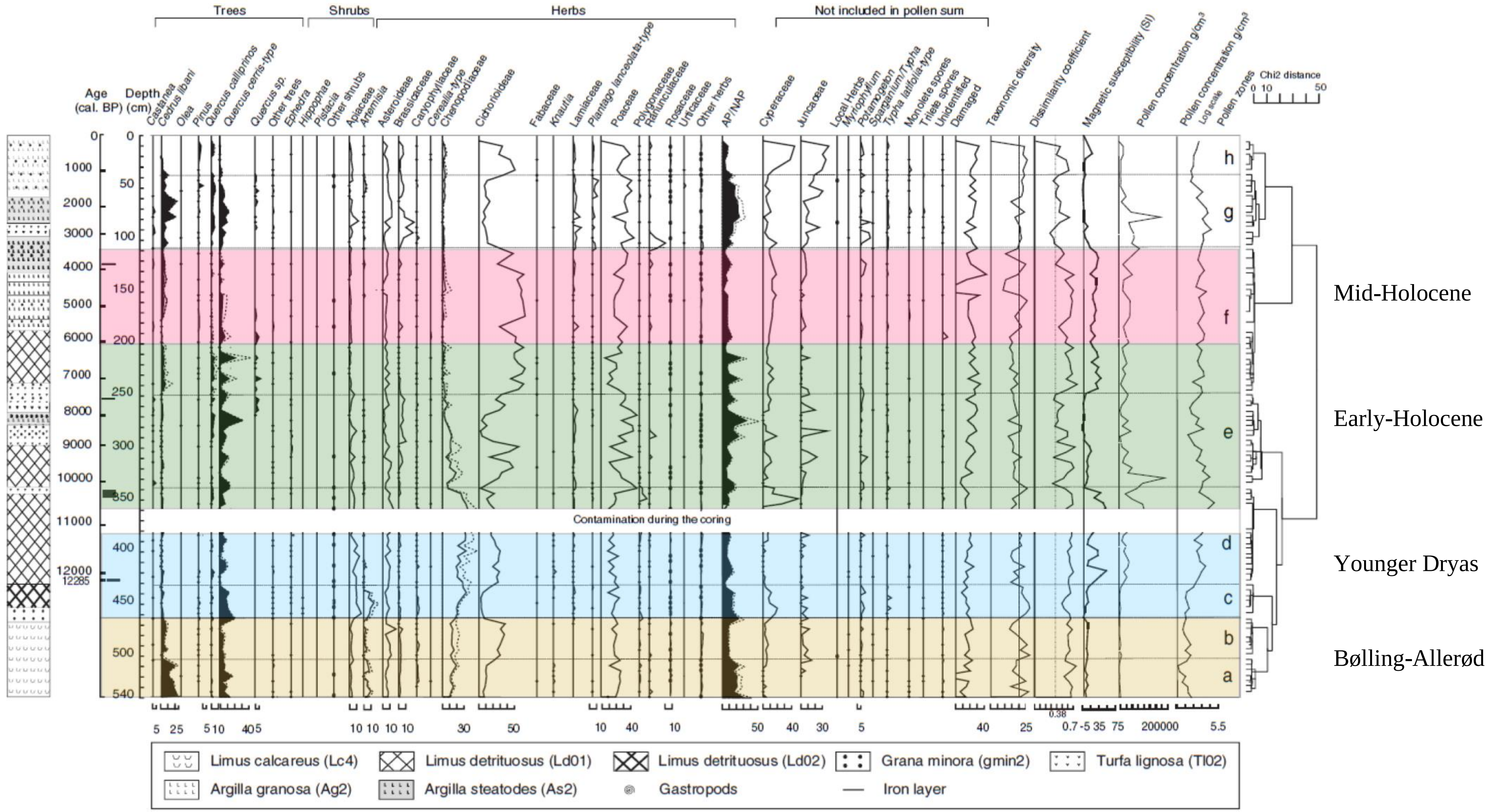

Figure 1.3 Pollen diagram from Aammiq Wetland (Hajar et al., 2008) that spans the Holocene and the late last deglaciation (~ 14 kyr cal. BP to present). Colour bands (yellow, Bølling-Allerød ; blue, Younger Dryas; green, Early-Holocene; pink, Mid-Holocene) highlight the different time periods and associated changes in vegetation community composition mentioned in the introductory text.

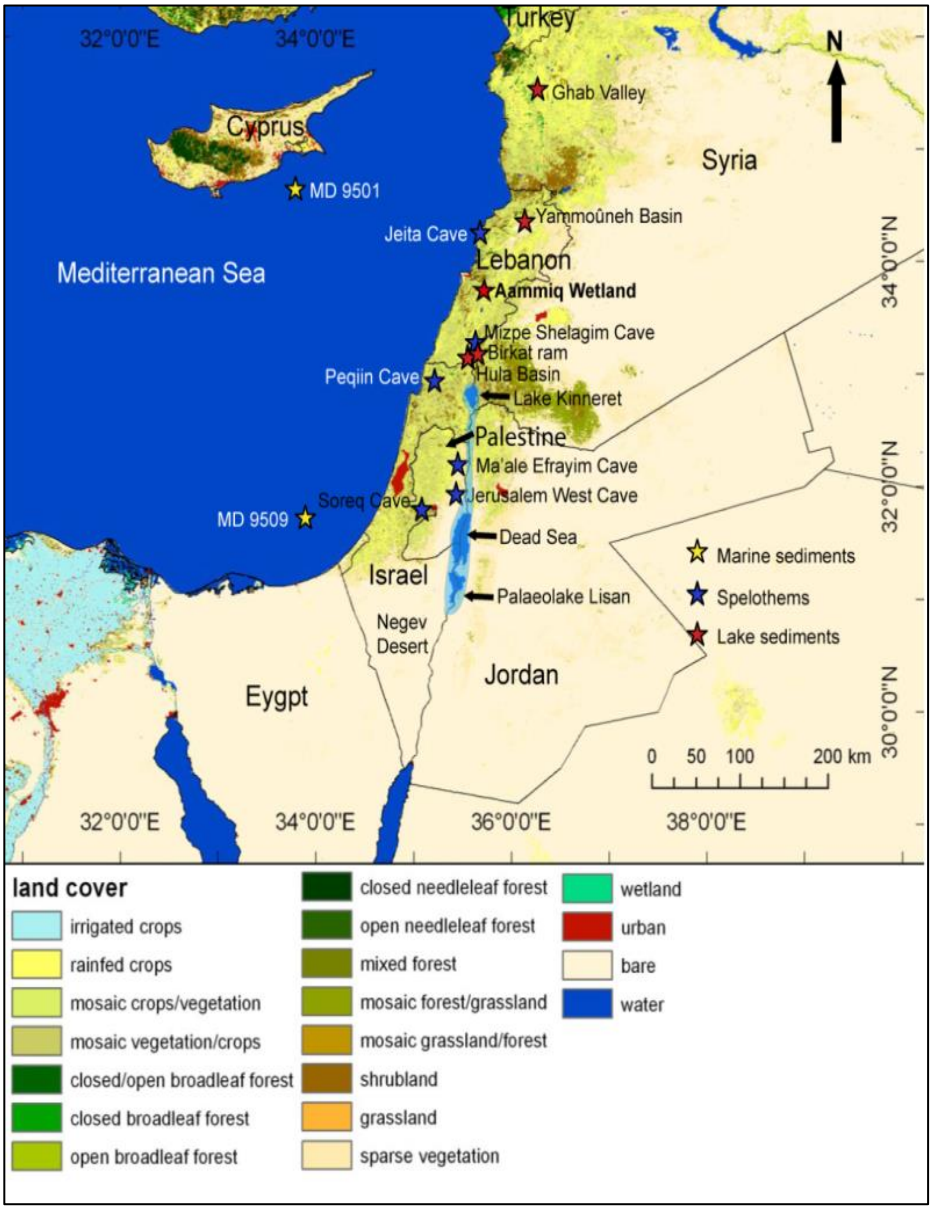

Figure 1.4 Land cover map of the Mediterranean Near East constructed using information from GLOBCOVER (Bircheron et al., 2009) also showing location of all study sites mentioned in introductory text.

± 0.13 kyr cal. BP (Fontugne et al., 1994), in response to pluvial conditions across the Mediterranean Basin (Kallel et al., 2004) and neighbouring continents (Rohling and Hilgen, 1991, Rohling and De Rijk, 1999). Temperature estimates based on speleothem data from Soreq Cave indicate that average temperatures in the southern Levant at this time were approximately 16 °C (McGarry et al., 2004), slightly higher than today (Affek et al., 2008).

The high levels of rainfall during the early-Holocene appear to have facilitated the growth of extensive C_3 vegetation (grasses and dicots) as far south as the northern Negev Desert, Israel (Fig.1.4) (Goodfriend, 1999). In other regions of the Levant, including the Ghab Valley, Syria (Niklewski and van Zeist, 1970, Yasuda et al., 2000) and the Hula Basin, Israel (Baruch and Bottema, 1999) (Fig.1.4) oak forests developed in the early-Holocene. However, the dating of these expansions of forest is problematic because the sediments from these sites suffer from large ‘reservoir effects’, which can lead to dating errors of several thousand years (Meadows, 2005).

From 6 kyr cal. BP a reduction in oak forest is observed in the Aammiq region that is possibly related to the onset of regional drying (Hajar et al., 2008). Palaeoclimatic data from speleothems at Jeita Cave (Verheyden et al., 2008), Soreq and Peqiin Caves, Israel (Fig. 1.4) (Bar-Matthews et al., 2003) and lake sediments from Yammoûneh Basin (Develle et al., 2010) indicate that at this time, there was a decrease in rainfall and a progressive rise in temperatures regionally. At approximately the same time, the decline of the coniferous forests in the Ghab Valley (Yasuda et al., 2000), increase in *Olea* in the Hula Basin (Baruch and Bottema, 1999) and at Birkat Ram, Golan Heights (Fig. 1.4) (Neumann et al., 2007) are all attributed to anthropogenic disturbance on the vegetation.

Time period (ka BP)	**Climate**			**Vegetation composition**		
	Regional	**Northern Lebanon**	**Southern Lebanon**	**Regional**	**Northern Lebanon**	**Southern Lebanon**
Mid-Holocene (~ 6-4)	Decreased rainfall and increased temperatures	Decreased rainfall and increased temperatures	Decreased rainfall and increased temperatures	Reduction in forest	Reduction in forest	Reduction in forest
Early-Holocene (~ 11-7)	Very wet and warm	Very wet and warm	Wet and warm	Oak forests	Oak forest	Open oak forest

Table 1.1 Summary of the climate and vegetation composition at the regional scale (Levant) and in southern and northern Lebanon during the early and mid-Holocene.

Last deglaciation (~ 18 to ~ 11ka BP)

The last deglaciation (~ 18 to ~11 ka BP), the transitional period between the end of the last glacial and the beginning of the current interstadial (Abrantes et al., 2012) was a period of rapid and extensive (Clark et al., 2012) climate warming (Broecker et al., 1988, Lenton et al., 2012). In Lebanon at Yammoûneh Basin, after ~ 16 ka BP the post glacial warming period was associated with an increase in moisture availability (Table 1.2) from melting glaciers and direct precipitation (Gasse et al., 2011). A similar increase in temperatures and precipitation at 17 ka BP was indicated by the isotope data from Soreq Cave (Bar-Matthews et al., 1999). The reduction in $\delta^{18}O$ values in *Globigerinodes ruber* foraminifera from marine cores 9501 (north eastern) and 9509 in the south eastern Levantine Basin in the EM (Fig. 1.4) after 15 ka BP (Almogi-Labin et al., 2009) is linked with a rise in sea surface temperatures (SST).

These warm and wet conditions favoured the expansion of oak forests (during the Bølling-Allerød warm interstadial 14.7 to 12.7 kyr cal. BP) in southern (Hajar et al., 2008) and northern Lebanon (Gasse et al., 2011), Hula Basin, Israel (van Zeist and Bottema, 1982, van Zeist and Bottema, 1991, Baruch and Bottema, 1999) and the Ghab Valley, Syria (Yasuda et al., 2000). The pollen data recovered from marine core 9509 for approximately the same time period (14.6-12.3 ka BP), also shows a significant increase in arboreal pollen (AP) levels (Langgut et al., 2011). The enhanced levels of rainfall in Lebanon at this time facilitated the development of a shallow lake in the Bekaa Valley at Aammiq (Hajar et al., 2008).

During the Younger Dryas (YD) event (~ 12.7-11.5 kyr cal. BP) that led to the abrupt return of glacial climate at high northern latitudes (Alley et al., 1993) the lake at Aammiq dried out and upland forests were reduced to small populations of cedar and deciduous oak within an open grassland mosaic dominated by steppic herbs (Hajar et al., 2008). This decline in AP and increase in steppic herbs during the YD was also recorded in marine core 9509, that are interpreted as indicating the existence of cold dry stadial conditions (Langgut et al., 2011). At Soreq Cave an increase in speleothem $\delta^{18}O$ values from 13.2-11.4 ka BP (Bar-Matthews et al., 1999) that appear to correlate in time to the YD event are inferred to indicate the onset of colder and dryer climatic conditions. However, some factor or combination of factors must have moderated this climatic event in northern Lebanon, as there is no indication of the YD event in the lake sediments recovered from Yammoûneh Basin (Develle et al., 2011, Gasse et al., 2011).

Time period (ka BP)		**Climate**			**Vegetation composition**		
		Regional	**Northern Lebanon**	**Southern Lebanon**	**Regional**	**Northern Lebanon**	**Southern Lebanon**
Last deglacial period (~ 18-11.5)	Bólling-Allerød (~ 16-13)	Increasing temperatures and precipitation	Increasing temperatures and humid conditions	Increasing temperatures and precipitation	Expansion of oak forests	Expansion of oak forest	Oak and cedar forest (~ 14-13)
	Younger Dryas (~ 13-11.5)	Cold and dry	Warm and wet?	Low precipitation and cold	Steppe	Oak forest	Grassland

Table 1.2 Summary of the climate and vegetation composition at the regional scale (Levant) and in southern and northern Lebanon during the last deglaciation.

Last glacial maximum MIS 2 (~ 18 to ~ 26 ka BP)

The last glacial maximum (LGM), the interval at the end of the last glacial period when ice sheets reached their maximum extent in northern latitudes (Mix et al., 2001), was associated with global temperatures and sea levels that were significantly lower than today (Yokoyama et al., 2000). At Yammoûneh Basin, Lebanon these low temperatures were concurrent with minimum values of arboreal and high non-arboreal pollen abundances that represent the presence of steppe like vegetation (Develle et al., 2011, Gasse et al., 2011). In conjunction with the low AP observed at Yammoûneh, the increased influx of detrital limestone gravel has led to the interpretation that the landscape was almost devoid of vegetation (Gasse et al., 2011) and that water availability in northern Lebanon had reached its minimum (Table 1.3) for the entire glacial period (Gasse et al., 2011). Estimates of temperature reconstructions based on $\delta^{18}O$ of lake sediments also indicate that temperatures declined at the LGM by as much as 10 °C lower than present day (Develle et al., 2010, Gasse et al., 2011) in good agreement with regional terrestrial and marine temperature reconstructions. At Soreq Cave stable isotopes of hydrogen (δD) from speleothem fluid inclusions (McGarry et al., 2004) and SST reconstructions based on alkelone measurements from marine core 9501 (Almogi-Labin et al., 2009) also show that terrestrial and SST temperatures respectively had declined by as much as 10 °C lower than today.

Time period (ka BP)	**Climate**			**Vegetation composition**		
	Regional	**Northern Lebanon**	**Southern Lebanon**	**Regional**	**Northern Lebanon**	**Southern Lebanon**
Last glacial maximum MIS 2 (~ 18-26 ka BP)	Low temperatures and precipitation?	Low temperatures and water availability	**?**	Steppe	Steppe	**?**

Table 1.3 Summary of the climate and vegetation composition at the regional scale (Levant) and in southern and northern Lebanon during the last glacial maximum.

Last glacial period MIS 3 to MIS 5d (~ 24 to ~110 ka BP)

As described above the LGM was the coldest interval of the whole glacial period. The 90,000 years of glacial conditions leading up to the LGM in northern Lebanon demonstrated significant climatic variability (Table 1.4) mapped upon a trajectory of declining temperatures and moisture availability (Develle et al., 2011, Gasse et al., 2011). The drying trend was punctuated by four humid episodes at ~ 105 to ~ 94 ka BP, ~ 80 to ~ 73 ka BP, ~ 65 to ~ 60 ka BP and ~ 45 to ~ 40 ka BP (Gasse et al., 2011). Evidence for this comes from three sources; sediment analysis, oxygen stable isotopes and pollen (Fig. 1.5) of lake sediments from Yammoûneh Basin. High concentrations of authigenic carbonates (Develle et al., 2011) reflect humid conditions because these carbonates only form in solution, and thus standing water (Develle et al., 2011). The decrease in the $\delta^{18}O$ values of ostracod valves are also interpreted as indicating wetter conditions (Develle et al., 2011, Gasse et al., 2011), and the relatively high percentages of tree pollen (oak and pine), that are taken to indicate the presence of forested landscapes, associated with an increase in moisture availability (Gasse et al., 2011). Between these humid episodes forest declined and the abundance of steppic herbs increased (Gasse et al., 2011).

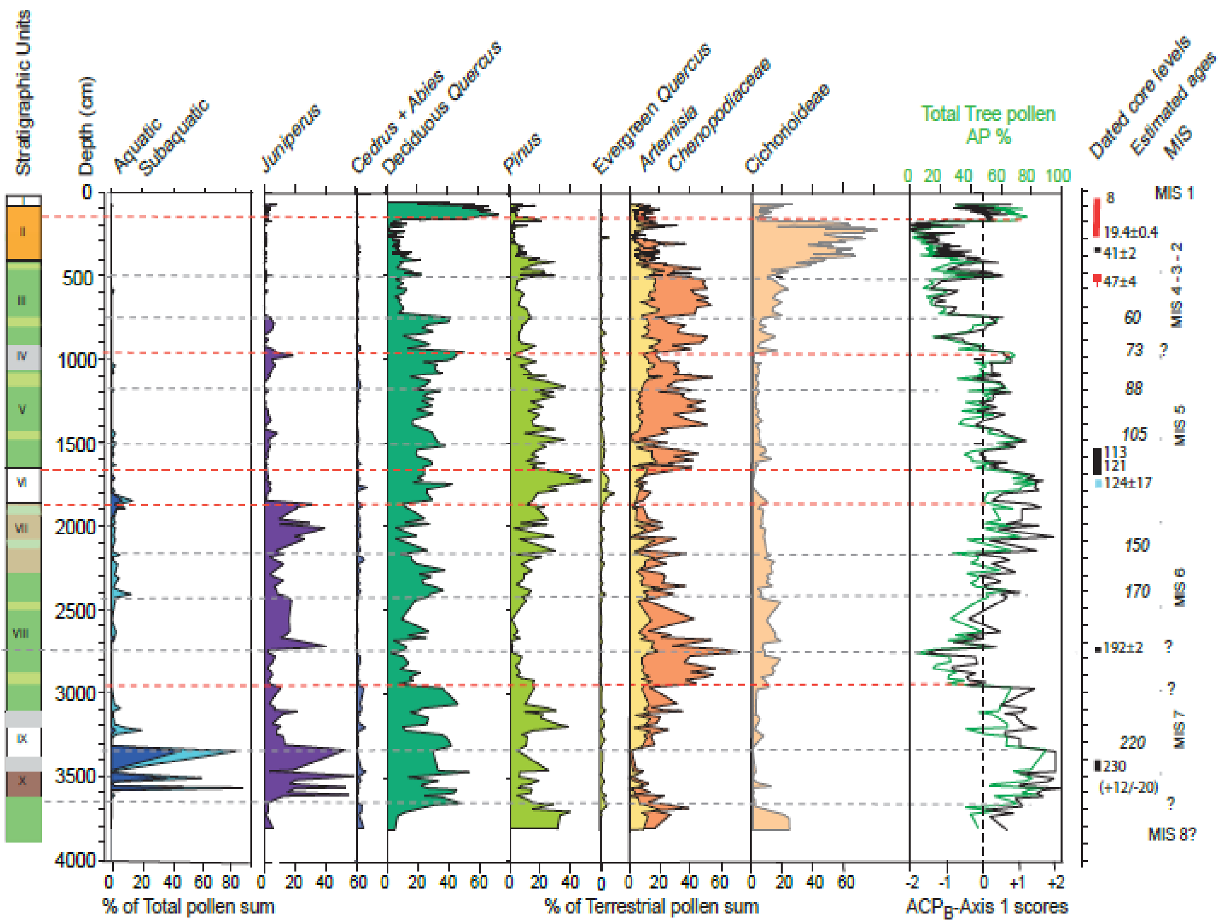

Figure 1.5 Pollen diagram from Yammoûneh Basin, northern Lebanon (Gasse et al., 2011) that spans the last 250,000 years. Blue shaded area represents aquatic and subaquatic species; purple, *Juniperus*; dark green, sum of *Cedrus libani*, *Abies* and deciduous *Quercus*; light green, *Pinus*; lime green, evergreen *Quercus*; yellow, *Artemisia*; orange, Chenopodiaceae; peach, Cichorioideae. The total percentage of tree pollen is displayed as the green curve on the right of the diagram.

The high levels of moisture availability recorded in Lebanon during MIS 5c (~ 105 to ~ 92 ka BP) and MIS 5a (~ 85 to ~ 70 ka BP), are coincident respectively with the deposition of a sapropels S4 and S3 in the Mediterranean Basin (Rossignol-Strick et al., 1981, Bar-Matthews et al., 2000). Similar humid events in the Levant were also observed in speleothems at Mizpe Shelagim Cave located on the slopes of Mount Hermon Golan Heights (Fig. 1.4) between 56 and 51 ka BP (Ayalon et al., 2013). The enhancement of rainfall during this interval is also observed at Soreq Cave (Bar-Matthews et al., 2000) and in marine core 9509, where a slight increase in AP is interpreted as indicating an increase in humidity (Langgut et al., 2011). In the Golan Heights this was followed by subsequent humid events at ~ 49, ~ 42 and ~ 36 ka BP (Ayalon et al., 2013). After ~ 40 ka BP there is no evidence of further humid periods until after the LGM in Lebanon (Develle et al., 2011, Gasse et al., 2011). The decline in either direct precipitation (Gasse et al., 2011) or effective precipitation (Develle et al., 2011) as a majority of water was contained in mountain glaciers, snow and permafrost soils on the mountain slopes favoured the presence of steppe like vegetation (Gasse et al., 2011) and resulted in the increased influx of detrital carbonates and gravels into Yammoûneh Basin (Develle et al., 2011).

While pollen evidence at Yammoûneh suggests arid conditions, in contrast reconstructed lake levels of Palaeolake Lisan (Fig. 1.4) to the south in the Dead Sea Basin (Israel/ Palestine) show increasing lake levels (Stein et al., 1997, Abed and Yaghan, 2000, Bartov et al., 2002, Bartov et al., 2003). It is suggested that this is due to higher direct precipitation input (Enzel et al., 2003) and therefore the presence of a north-south Levant rainfall gradient during the last glacial period that reached its maximum difference during MIS 3 and MIS 2 (Develle et al., 2011).

Time period (ka BP)	Climate			Vegetation composition		
	Regional	Northern Lebanon	Southern Lebanon	Regional	Northern Lebanon	Southern Lebanon
Last Glacial MIS 3 (~ 60-26)	Cold and dry to cold and wet	Cold and dry to cold and humid	?	Steppe with oak forest expansion (~ 56-51)	Steppe-with open oak forest expansion (~ 45-40)	?
MIS 4 (~ 70-60)	Dry and cold?	Cold and dry to cold and wet	?	Steppe	Steppe-with oak forest expansion (~ 65-60)	?
MIS 5a (~ 85-70)	Wet	Wet and warmer relative to MIS 5b	?	Oak forest	Oak and pine forest	?
MIS 5b (~ 93-85)	Dry and cool	Dry and colder relative to MIS 5c	?	?	Steppe	?
MIS 5c (~ 105-93)	Wet	Wet and warmer Relative to MIS 5d	?	?	Oak and pine forest	?
MIS 5d (~ 110-105)	Dry	Dry	?	?	Steppe	?

Table 1.4 Summary of the climate and vegetation composition at the regional scale (Levant) and in southern and northern Lebanon during the last glacial period prior to the last glacial maximum (~ 24 to ~ 110 ka BP).

In summary, all of the available climatic data from southern and northern Lebanon, and the wider Levant shows that the climate varied significantly through time from the start of the last glacial period up to the present day (Table 1.5). Placing the climatic records from Lebanon into a regional context suggests that the climate in Lebanon primarily followed a similar trajectory through time to that observed regionally but also displays some distinct differences. In response to these temporal changes in the climate regime, it is known that the composition of the vegetation community in northern Lebanon (~ 110 ka BP to present) and southern Lebanon (~ 14 ka BP to present) oscillated between open oak forests and steppe (Table 1.5).

Time period (ka BP)		Climate			Vegetation		
		Regional	**Northern Lebanon**	**Southern Lebanon**	**Regional**	**Northern Lebanon**	**Southern Lebanon**
Mid-Holocene (~ 6-4)		Decreased rainfall and increased temperatures	Decreased rainfall and increased temperatures	Decreased rainfall and increased temperatures	Reduction in forest	Reduction in forest	Reduction in forest
Early-Holocene (~ 11-7)		Very wet and warm	Very wet and warm	Wet and warm	Oak forests	Oak forest	Open oak forest
Last deglacial (~ 18-11.5)	Bólling-Allerød (~ 16-13)	Increasing temperatures and precipitation	Increasing temperatures and humid conditions	Increasing Temperatures and precipitation	Expansion of oak forests	Expansion of oak forest	Oak and cedar forest (~ 14-13)
	Younger Dryas (~ 13-11.5)	Cold and dry	Warm and wet?	Low precipitation and cold	Steppe	Oak forest	Grassland
Last glacial maximum (~ 26-18)		Low temperatures and precipitation?	Low temperatures and water availability	?	Steppe	Steppe	?
Last Glacial MIS 3 (~ 60-26)		Cold and dry to cold and wet	Cold and dry to cold and humid	?	Steppe with oak forest expansion ~ 56-51	Steppe with open oak forest expansion ~ 45-40	?
MIS 4 (~ 70-60)		Dry and cold?	Cold and dry to cold and wet	?	Steppe	Steppe- with oak forest expansion ~ 65-60	?
MIS 5a (~ 85-70)		Wet	Wet and warmer relative to MIS 5b	?	Oak forest	Oak and pine forest	?
MIS 5b (~ 93-85)		Dry and cooler	Dry and colder relative to MIS 5c	?	?	Steppe	?
MIS 5c (~ 105-93)		Wet	Wet and warmer relative to MIS 5d	?	?	Oak and pine forest	?
MIS 5d (~ 110-105)		Dry	Dry	?	?	Steppe	?

Table 1.5 Shows the consolidation of Tables 1.1 to 1.4 summarising the changes in climate and vegetation composition at the regional scale (Levant) and in southern and northern Lebanon from the start of the last glacial period to the Mid-Holocene (~ 110 to ~ 6 ka BP).

1.2 Aims

Above, I have reviewed the current state of knowledge regarding the environmental history of the Aammiq region, southern Lebanon and placed it within a regional context of known environmental events in Lebanon, the Levant (Israel, Palestine, Syria, Jordan) and the eastern Mediterranean Basin. What remains unknown however is how the climate in the Aammiq region of southern Lebanon altered during the last glacial cycle prior to the onset of the deglaciation. Additionally, we do not how the composition of the vegetation in this region of Lebanon altered through time in response to either changes in climate or other disturbance mechanisms. Also, at present very little is known about how climate change affected landscape processes. The aims of this thesis are thus: Firstly, to reconstruct the climate of the Aammiq region during the last glacial period. Secondly, examine how the composition of the vegetation communities in this region altered in response to changes in climate. Thirdly, to show how climate change affected other environmental processes and in turn how these affected plant communities and the structure of the wetland. These aims are addressed by considering four major research questions:

1. *How did the climatic conditions of the last glacial period vary in southern Lebanon? What was the effect of this climatic variability on landscape processes and the wetland in the Bekaa Valley?*
2. *What was the plant community response to glacial conditions at the last interglacial-glacial transition?*
3. *How did climatic variability throughout the last glacial period alter the structure and composition of plant communities in the Bekaa Valley and on the Lebanon Mountains?*

4. *What was the frequency of fire activity in and around Aammiq Wetland during the last glacial period? How did this relate to climatic variability? And what was the impact of these disturbances on the vegetation community?*

In order address these questions, a multi-proxy palaeoenvironmental study was conducted on lake sediments recovered from Aammiq Wetland in the southern Bekaa Valley, Lebanon. The results obtained have been compiled into four papers submitted for peer review that are outlined below and detailed in **Chapters 4-7**. Each of these chapters (papers) includes its own introduction to the topic being addressed, description of the study site, methods used, results, discussion and conclusion. Please note that these papers are based on the analysis of sediments (employing different proxies) from the same sediment core there is therefore inherently some repetition in these papers in the description of the study site and dating methodologies.

1.3 Outline of thesis

In **Chapter 2**, a detailed description of the study site (Aammiq Wetland) is provided and the reasons behind the core-site selection are outlined.

In **Chapter 3**, the methods, laboratory techniques and statistical analyses used in this research are explained. Additionally, the factors constraining analysis of the entire length of the lake sediment core extracted from Aammiq Wetland are explained and which sections of the sediment core are used for which analyses in each chapter are detailed. Also, the age depth-model output that is fundamental to this study is presented along with the preliminary results of the geochemical analysis detailed in **Chapter 4**, the vegetation reconstructions that are presented in **Chapters 5** and **6** and the micro and macrocharcoal analysis detailed in **Chapter 7**.

Chapter 4: Landscape erosion, karstic activity and the development of a wetland in the southern Bekaa Valley, Lebanon during the last glacial period.

Darren Jeffers and Kathy Willis

Submitted to Paleolimnology 6th October 2013 (**in revision**)

In this paper we present the results of a sedimentological analysis of a sedimentary sequence recovered from the Bekaa Valley, Lebanon that spans the end of the last interglacial and the last glacial period (~ 112 to ~ 35 ka BP). The primary aims of this paper are firstly to use both the sediments geochemical and physical properties to assess how erosion and weathering of the montane landscape varied through the last glacial period and secondly to use these data to infer how local climate varied. In addition, to ascertain what was the effect of climate change and landscape processes on the water body in the Bekaa Valley?

Chapter 5: Vegetation response to climate change during the last interglacial-last glacial transition in the southern Bekaa Valley, Lebanon.

Darren Jeffers and Kathy Willis

In press Palynology 2014

During the last interglacial period extensive forests covered the slopes of the Lebanon Mountains. This paper investigates the response of these forests to the abrupt and extensive climatic change at the last interglacial-last glacial transition (MIS 5e to MIS 5d, ~ 112 to ~ 104 ka BP). What was the effect of this abrupt climate shift on the terrestrial and aquatic vegetation? In particular, how did the emblematic species cedar of Lebanon respond to the change in moisture availability that climate scenarios predict will occur in the future?

Chapter 6: Eastern Mediterranean montane vegetation variability through the last glacial cycle-a record from the Bekaa Valley Lebanon.

Darren Jeffers and Kathy Willis

Submitted to Palaeogeography, Palaeoclimatology, Palaeoecology 4th October 2013 (**in review**)

The last glacial period in the Mediterranean was comprised of stadial (cold and dry) and interstadial (warmer and wetter) intervals. The duration of these alternate climatic states varied from less than a 1000 years to the multi-millennial scale and exhibited significant variation in their magnitude. This study uses a 70,000 year time series of pollen data to investigate the relationship between climate change and vegetation variability through the last glacial cycle (~ 104 to ~ 35 ka BP). Were changes in the composition of the plant community concurrent and associated with known orbital, millennial or centennial-scale global climatic events? And furthermore, were forests and in particular populations of cedar present in some form throughout the last glacial period?

Chapter 7: Orbital and millennial-scale forcing of fire activity in Lebanon during the last glacial period

Darren Jeffers and Kathy Willis

Submitted to Quaternary Research 26th September 2013 (**in revision**)

Wildfires are a naturally occurring phenomenon in the Mediterranean Basin that are strongly associated with the hot and dry climatic conditions during the summer. This paper explores the relationship between fire activity, vegetation variability and climate change during the last glacial period (~ 104 to ~ 35 ka BP). The primary aim is to ascertain whether there are any trends in peak fire activity and assess if any association exists between these peaks and changes in orbital parameters or global climate events at the millennial-scale. Additionally, it also investigates the potential role of fire as a driver of ecological change and whether fire/vegetation feedback mechanisms exist and their function in controlling fire frequency and intensity.

Chapter 8: Conclusion

At the conclusion of this thesis I summarise and synthesis the major findings of the four papers outlined above. In the last part of the conclusion I describe future research that will build on the research described within this thesis.

1.4 Summary

This thesis presents a detailed investigation into the key processes of abiotic and biotic environmental change in the Aammiq region of Lebanon during the last glacial period. This is achieved by employing a wide range of laboratory techniques to extract and analyse a variety of physical and chemical proxies contained within the lake sediments recovered from Aammiq Wetland. The findings are compared and contrasted to previous research from within Lebanon and the Levatine region of the Near East to place into its regional context the observed environmental change at Aammiq. Furthermore, the effects of changes in Earth's orbital parameters and global climatic change events on the environment in the Aammiq region are considered.

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Chapter 2: study site

2.1 Location

Aammiq Wetland is situated in the southern Bekaa Valley, Lebanon (33°43'50"N, 35°47'10"E) at 860 meters above sea level (m.a.s.l) (Fig. 2.1). The wetland has a high international conservation status; it is designated as a RAMSAR site, UNESCO Man and Biosphere Reserve, and an Important Bird Area (IBA). It lies on the western side of the valley plain at the foot of the eastern slopes of the Lebanon Mountain range. The contemporary wetland is a small remnant of far more extensive wetlands that existed until the 1940’s, as indicated by historical maps (Walley, 1996). Local inhabitants confirm that the wetland extended at least 30 km north of its current boundary (Storey, 2007). The extent of the wetland was progressively reduced as the wetlands were drained to enable the expansion of agricultural production (Walley, 1996, Storey, 2007). Today almost the entire flat fertile valley plain is given over to the cultivation of crops, which include citrus fruits, grapes, vegetables and cereals (FAO, 2008).

2.2 Core site selection

The core site location within the wetland (33°43'51.47"N, 35°47'08.80"E) is shown in figure 2.2. Also shown is the location of the Hajar et al., (2008) core site (33°43'57.12"N, 35°47'07.52"E) that provided a sediment core extending back in time through the Holocene to the last deglaciation (~14 ka BP to present). The core site for this study was chosen in the knowledge that during the wet season (January-March) the location is comprised of an open water pool surrounded by reed beds of *Phragmites australis* (common reed) and *Scirpus* (bul-rush). Coring was conducted during the autumn (November, 2008) when the wetland was dry to provide a firm drilling platform that dispensed with the need for either a boat or a floating platform.

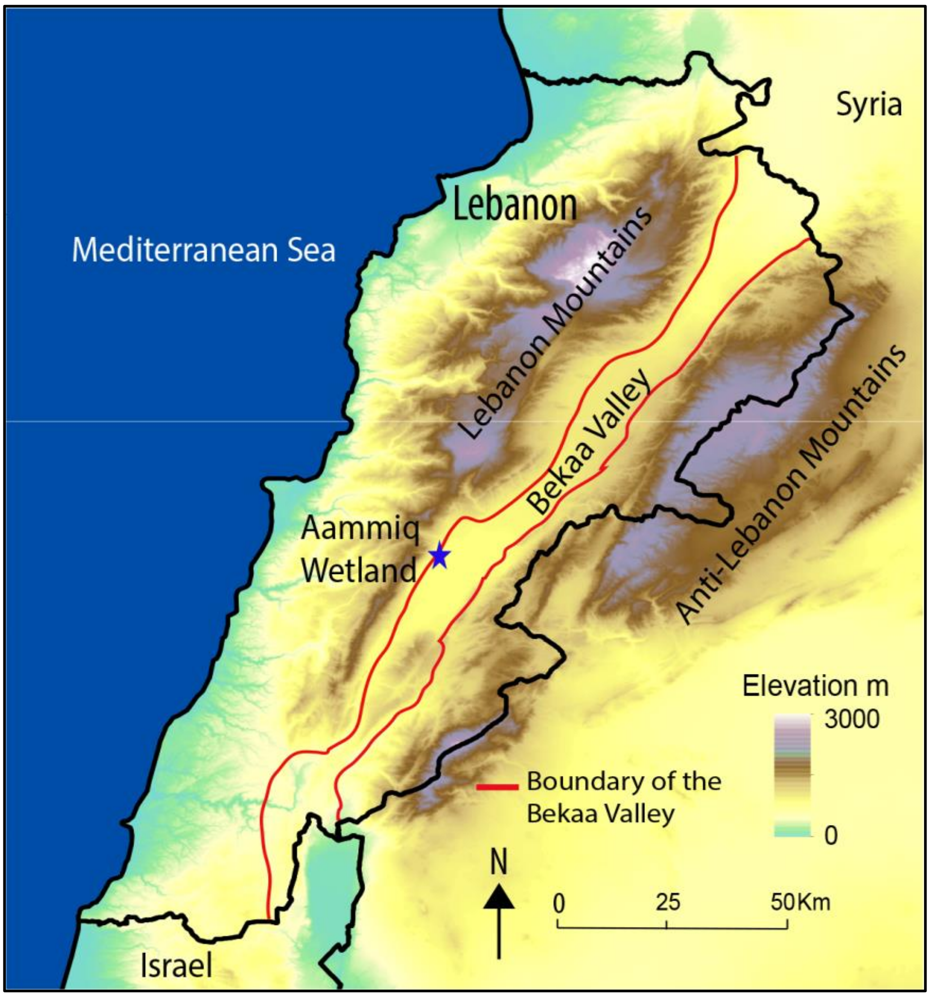

Figure 2.1 Topographic map of Lebanon showing location of Aammiq Wetland within the Bekaa Valley.

2.3 Climate

The climate is Mediterranean in type with warm dry summers and cool wet winters. Average temperatures in the southern Bekaa Valley range from 2 °C in winter to 34 °C in summer (Ministry of Environment/Med Wet Coast Project, 2004). Average annual precipitation (1954-2002 long term average, as recorded at Tal Amara meteorological station, 30 km north of Aammiq) is only 600 mm yr^{-1} (Fig. 2.3) (Karam et al., 2007) in comparison to Lebanon's coastal areas, which receive an average of between 700-1000 mm yr^{-1} (Asmar, 2011). This dissimilarity is due to the rainshadow effect of the Lebanon Mountains which

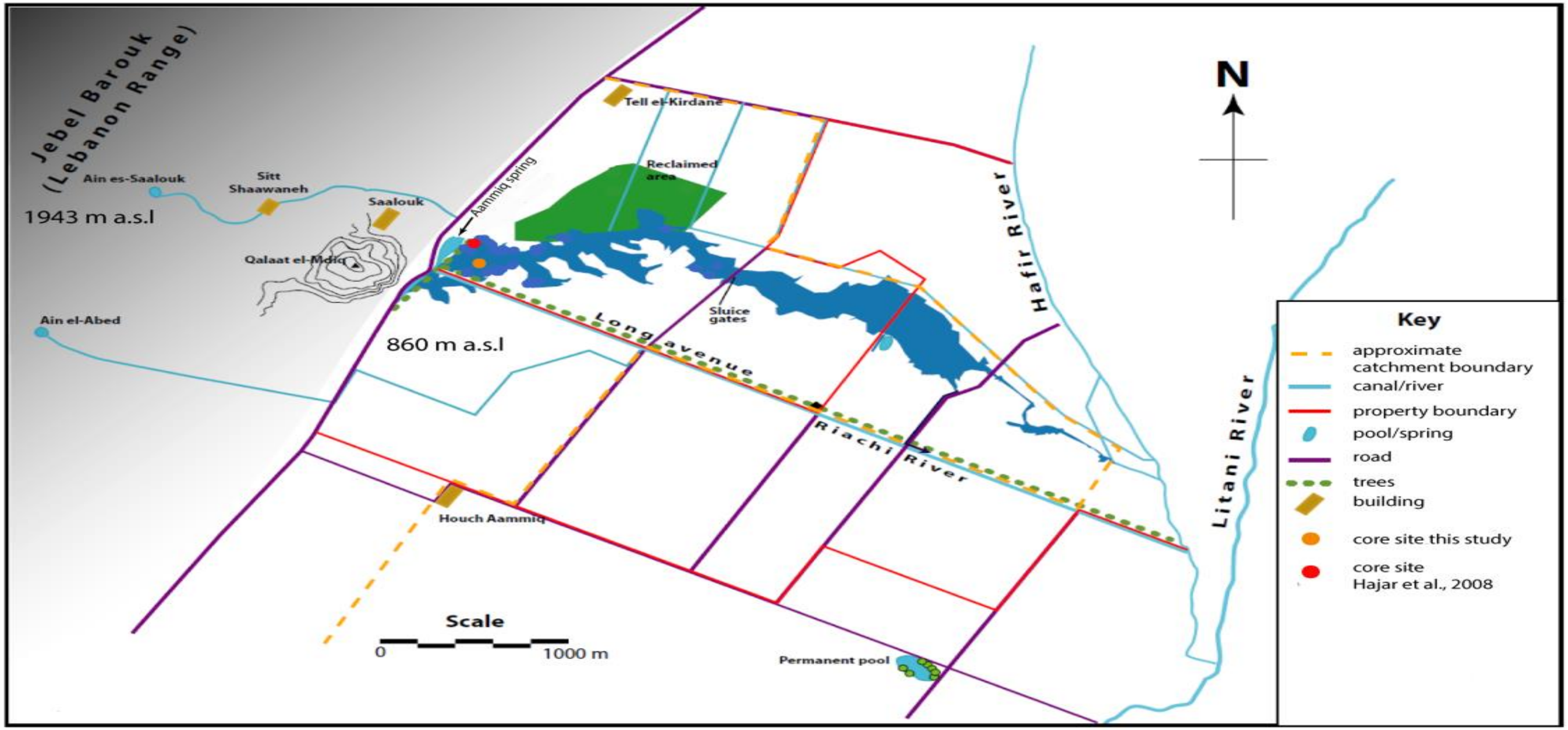

Figure 2.2 Map of the Aammiq region (Adapted from: Storey, 2007) showing the approximate extent of the wetland during the wet season (January-March), the location of Aammiq spring that provides a majority of the water input into the wetland and the wetlands proximity to and altitude in comparison to the Lebanon Mountains. Also shown is the location of the core site of this study and its position relative to that of the Hajar et al. (2008) core site.

reduces the moisture content of airflows that pass from the Mediterranean Sea over to the Bekaa Valley in the east (Sene et al., 1999). Precipitation occurs on 80-90 days a year, mainly between November and April (Storey, 2007). In dry years precipitation can be less than 50 percent of the long term average (Storey, 2007). Due to the high summer temperatures, evaporation can exceed 70% of precipitation in this region of the interior of Lebanon (NCFRS, 1998). On the higher elevations of the Lebanon Mountains average annual precipitation can exceed 1600 mm yr^{-1} (Asmar, 2011), with a majority of the precipitation falling as snow (FAO, 2008). On the higher mountain slopes average annual temperature of the coldest months (December, January, February) is below 0 °C and the warmest month (July) more than 30 °C (Gasse et al., 2011).

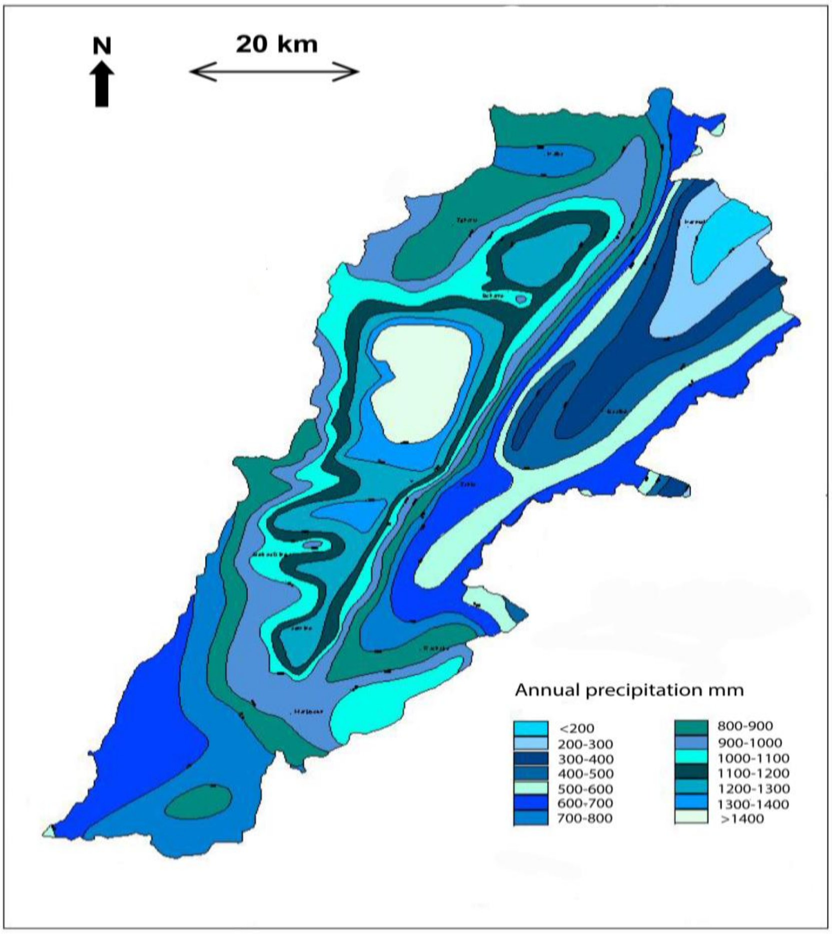

Figure 2.3 Isohyet map of Lebanon showing average annual precipitation (Source: Plassad, 1971).

2.4 Geology and geomorphology

The major geological structures of this region are related to the NE-SW aligned Yammoûneh Fault System (Fig. 2.4), which forms part of the Dead Sea Transform Fault that extends from the NW tip of the Red Sea to the Taurus Mountains of Anatolia, and is the result of the intersection of the Arabian and African plates (Walley, 1996, Walley, 1998).This fault line is responsible for the abrupt transition in topography from the Lebanon Mountains

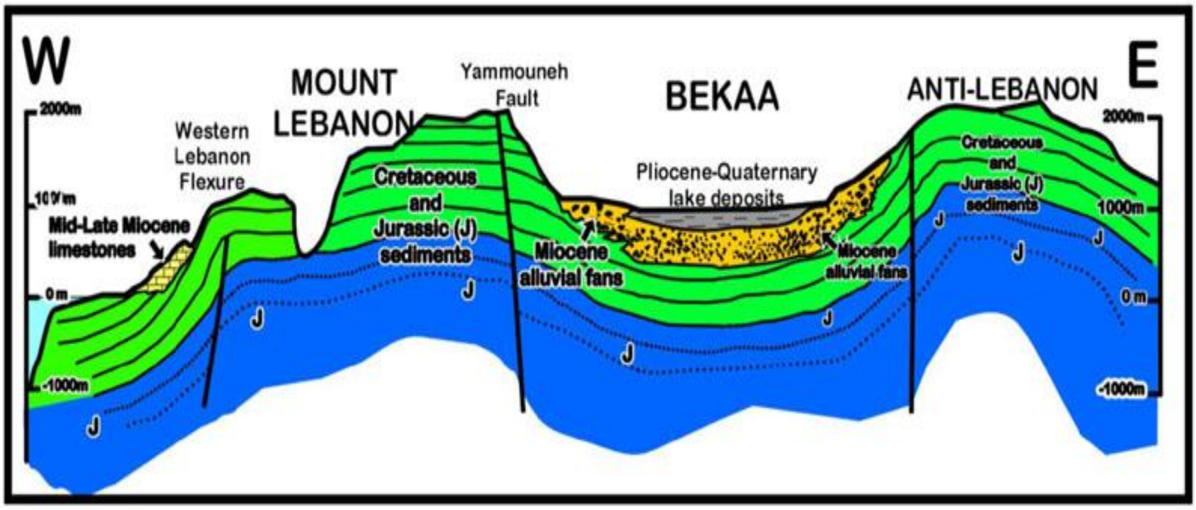

Figure 2.4 Cross sectional diagram of Lebanon showing main geological features (Source: Walley, 1998).

to the NE and the Bekaa Valley to the SW (Walley, 1996, Walley, 1998). The Lebanon Mountain range (max. elevation Qurnat as Sawda 3088 m) that runs approximately parallel to the Mediterranean coastline is the result of plate deformation that began in the late-Cretaceous (Dubertret et al., 1970) ~100 to ~ 90 Ma BP and continued until the Oligocene (23 Ma) epoch (Walley, 1998). Parallel to this mountain range to the east of the valley are the Anti-Lebanon Mountains (highest peak Jabal el Shaykh 2814 m). The section of the Lebanon Mountains adjacent to the wetland (named Jebel Barouk at this point) attains an elevation 1943 m and is predominately comprised of Cretaceous and Jurassic limestones (Walley, 1996). The shallow soils on the higher slopes of the Lebanon Mountain are composed of lithic leptosols (Asmar, 2011) and inceptisols (Vogiatzakis, 2012), and in areas that have been subjected to significant human activity terric anthrosols (Asmar, 2011). Erosion of these mountainous soils in Lebanon is a serious problem due to the steep topography (1:2) and has been accelerated by historical deforestation and high grazing pressures (Asmar, 2011). The sediments contained in the alluvial fan (> 1.5 km in width) that descends the flank of the Lebanon Mountains that ends at the wetland's edge are comprised of a mixture of sandstone (deposited during the Miocene) and limestone clasts (Walley, 1998). In the valley the soils are mainly alluvial and colluvial in origin that take a variety of forms including calcaric cambisols, eutric vertisols, calcaric arenosols and eutric fluvisols (Darwish et al., 1999). The valley sediments underlying these soils are comprised mainly of alluvial deposits consisting of silts, clays, sand and gravel that were laid down during the Quaternary period (Walley, 1998), and are probably in excess of 1 km in depth in parts of the wetland (Walley, 1998).

2.5 Hydrology

Water input into the marsh occurs through a combination of direct precipitation and the inflow of spring water (Walley, 1996). Contemporary levels of precipitation are insufficient to support the persistence of the wetland (Walley, 1996). A far greater input of water derives from numerous springs in the area that are fed from karstified aquifers situated within the slopes of the Barouk Mountain (Walley, 1996, Storey, 2007), that are recharged from snowmelt and rainfall (Walley, 1996). The marsh is generally full of water from January to March, with water levels declining from May, and in a typical year the wetland is totally dry by late July until December (Storey, 2007). The wetland can temporarily expand in area up to 300 ha during periods of heavy rainfall and snow melt. The area that is continuously flooded, fed by the influx of spring water during the wet season, is generally limited to approximately 70 ha (Storey, 2007). Inter-annual variability in average annual precipitation in the Bekaa, leads to high variations in water levels and extent of the marsh (Storey, 2007).

The water source for Aammiq has been diverted over the last four decades for agricultural production, which has lowered the groundwater table by several meters (Walley, 1996). This process of drawdown has resulted in the creation of the ephemeral wetland that exists today.

2.6 Vegetation

The wetland today is a mosaic of different habitats, whose spatial configuration corresponds with different hydrological conditions. Open grassy habitats form outside the main wetland boundary during prolonged periods of heavy rain; these persist for only a few days or weeks (Storey, 2007). Within the wetland, reed beds are the most abundant habitat (Conroy and Hepper, 2005) and are comprised of emergent plants (e.g *Phragmites australis*, common reed, *Typha spp*., reed-mace and *Scirpus*, rushes). The reed beds are inter-dispersed by open water pools up to depths of 2-3 m that contain over 200 floral species of submerged (e.g *Potamogeton spp*., pondweed), floating plants (e.g *Lemna spp.*, duckweed) and *Ranunculus spp*. (water crowfoot) as well as algae *Chara spp*. (stonewort) (Conroy and Hepper, 2005).

The forests on the flank of the Lebanon Mountains are fragmented and largely open as a result of intensive timber extraction (Fady et al., 2008). The vegetation assemblages can be differentiated into three bioclimatic zones according to altitude. From the wetlands edge (860 m) to 1000 m a.s.l (Eu-Mediterranean zone), degraded *Quercus calliprinos* (Palestine oak) forest persists, whereas between 1000 and 1600 m a.s.l (Supra-Mediterranean) the slopes are scattered with *Quercus infectoria* (gall oak) and *Juniperus oxycedrus* (prickly juniper). Lastly above 1600 m a.s.l (Mountainous Mediterranean) stands of *Cedrus liban*i (cedar of Lebanon) are the predominant tree (Hajar et al., 2008). The cedar forests are primarily located in the Al-Shouf Cedar Reserve that incorporates the peak and upper slopes of the western flank of the southern section of the Lebanon Mountains. These forest remnants at all elevations are inter-dispersed by shrub vegetation dominated by the dwarf shrub *Sarcopoterium spinsum* (thorny burnet) that in arid and semi-arid regions of the Mediterranean is considered an indicator of land degradation (Ackermann et al., 2004).

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Chapter 3: Methodology and preliminary results

3.1 Sediment sampling

This study is based on a multi-proxy study of a 7 m sedimentary sequence collected from Aammiq Wetland, Lebanon (33°43'51.47"N, 35°47'08.80"E). The sediment core was extracted using a Russian corer that collected a semi-cylindrical shaped sample with a 5 cm radius. To facilitate the extraction of the sediment core from the wetland a steel A-frame fitted with a wire rope winch was constructed on-site (Fig. 3.1).

Figure 3.1 Photographs of the core site with Lebanon Mountains in the foreground, excavation of sediment around core site to facilitate extraction, and A-frame and winch.

3.2 Chronology

Age chronology for the core was developed using a combination of accelerator mass spectrometry (AMS) radiocarbon dating (2 bulk samples, 360 and 258 cm) conducted at ^{14}CHRONO Radiocarbon Laboratory, Queen’s University Belfast, Belfast, UK and optically stimulated luminescence dating (OSL) (5 quartz samples, 219, 278, 486, 544 and 690 cm) performed at the Oxford Luminescence Laboratory (OLD) at the University of Oxford,

Oxford, UK. A third radiocarbon date (318 cm) was omitted from the construction of the age-depth model as initially it failed to produce any date due to low graphite yield (Hoper, pers comm., 2012 a) and on being run for a second time produced an anomalously young date, that could reflect either sediment disturbance or modern contamination. The 5 quartz samples that were subject to OSL dating were the only samples that contained any quartz from the initial 30 sediment samples that were subjected to chemical treatment (detailed in **Chapters 4-7**) to isolate the quartz fraction of the sediment.

Radiocarbon dates from bulk sediments in a region where the bedrock is primarily comprised of limestone (Edgell, 1997, Walley, 1996) are purported to have been subject to the 'reservoir effect' of geological old carbon entering the aquatic system (Develle et al., 2010), thereby producing age estimates that are too old by up to several thousand years (Meadows, 2005). An attempt to circumnavigate this effect was made by extracting the terrestrial fossil pollen fraction from the sediment for radiocarbon dating following the protocol of Brown et al. (1989), thereby removing aquatic material from the dating process. Unfortunately the concentrations of pollen sent to ^{14}CHRONO and hence carbon levels were too low to produce any signal (Hoper, pers comm., 2012 b).

Therefore to determine the extent of any potential reservoir effect on radiocarbon dates obtained from bulk organic material the surface sediment from 0-1 cm depth was analysed by AMS at the Oxford Radiocarbon Accelerator Unit (ORAU), University of Oxford, Oxford, UK. The result of the radiocarbon dating is presented in table 3.1. Calibration of the ^{14}C age to years BP (cal. yr BP) was performed using OxCal version 4.2.3 (Bronk Ramsey, 2013) that used the post-bomb atmospheric curve 13 NH2 (Hua et al., 2013). The calibrated radiocarbon date is 1958-59 and is therefore negative or post-bomb and rich in ^{14}C to the extent that the sediment must have been exchanging with the atmosphere after 1950 AD. This suggests that in recent decades there has been very little if any old carbon from geological sources entering the wetland. Given these findings it was decided that it was reasonable not to apply an age offset when calibrating ^{14}C ages obtained from the wetland's bulk sediments.

Sample depth (cm)	Lab ID	Material	Fraction modern carbon (FMC)	Calibrated date (yr AD) to 2σ
0-1	OxA-29284	Bulk sediment	1.18926 ± 0.00346	1958-1959

Table 3.1 Radiocarbon dating results for bulk sediments from 0-1 cm from the Aammiq Wetland sedimentary sequence. Calibration was performed using OxCal version 4.2.3 (Bronk Ramsey, 2013) using the post-bomb atmospheric curve 13 NH2 (Hua et al., 2013).

Calibration of the bulk sediment (360 and 258 cm) radiocarbon dates (^{14}C ages) to years before present (cal. yr BP) without applying any offset was performed using CLAM's default northern hemisphere (atmospheric) calibration curve that uses IntCal09 (Reimer et al., 2009). The Age-depth model for the sedimentary sequence combining the calibrated radiocarbon dates and the OSL dates was then constructed using linear interpolation in the software program CLAM version 2.1 (Blauuw, 2010) run in R programming language (2010). The output of the age-depth model as a graph is shown in figure 3.2.

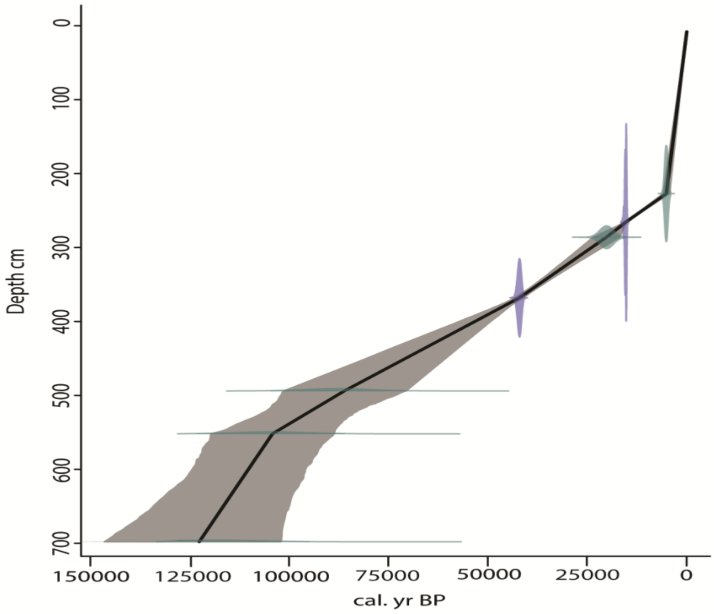

Figure 3.2 Age-depth graph produced in CLAM version 2.1 (Blaauw, 2010) using linear interpolation. Black line represents the 'Best fit' model and the grey shaded area 95% confidence limits. The probability distribution of the calibrated radiocarbon dates are shown as blue reduced plots, and the probability distribution of the OSL dates as green reduced plots.

3.3 Fossil pollen

Fossil pollen was extracted from the sediment to enable reconstruction of temporal variations in vegetation. Although the outer layer (exine) of pollen grains that is comprised of sporopollenin is very resistant to both physical and chemical processes (Moore et al., 1991, Bennett and Willis, 2001) it is susceptible to degradation due to oxidation (Bennett and Willis, 2001, Hopkins and McCarthy, 2002). Exposure to oxygen and subsequent degradation is a particular problem for the preservation of pollen in arid climatic regions (Horowitz, 1992) and previous work in the study region had indicated that this was indeed a potential problem resulting in low pollen contractions (Hajar et al., 2008). In an attempt to devise a pollen extraction method that would consistently result in the concentration of sufficient identifiable pollen (> 200 grains per-level) experimentation with multiple variations and combinations of "standard preparation" (e.g Faegri et al., 1989, Moore et al., 1991, Bennett and Willis, 2001), heavy liquid gravity separation (Smith, 1998, Nakagawa et al., 1998, Lentfer and Boyd, 2000) and sieving (Woosley, 1978, Horowitz, 1992) pollen extraction methods was therefore undertaken.

The pollen extraction method that produced the greatest concentrations of pollen was comprised of a series of steps from a "standard preparation" and sieving methodologies. This method omitted the application of the alkali (e.g NaOH-sodium hydroxide) used to remove humic acids and disaggregate the sediment, which causes deterioration of pollen grains already exposed to oxidation (Woosley, 1978) and the acetylation (acetolysis) process using sulfric acid (H_2SO_4) that can degrade and destroy pollen (Hesse and Waha, 1989, Charman, 1992). Instead the chemical treatment that was undertaken included 7% hydrochloric acid (HCl) to remove carbonates (90 °C for 30 minutes). As no base (alkali) was employed as a deflocculant, the sediment was vigorously stirred whist in HCl to initially separate the sediment fractions. The sediment was then treated in hydrofluoric acid (HF) 58-62% to remove silicates (90 °C for 90 minutes), 7% HCl (90 °C for 90 minutes) to remove colloidal silica and silicofluorides created by the HF reaction. The remaining sediment was washed to remove humic acids and clays using a fine sieving approach (10 μm) over a vacuum flask. This method minimised loss of pollen from already low concentrations by reducing decanting by 50% relative to a "standard preparation" pollen extraction method. The remaining fraction was then dehydrated using tert-butyl alchol (TBA). The remaining residue containing pollen was then mounted in silicone oil.

Pollen identification was conducted using microscopy at ×200, ×400 and ×1000 magnification. Pollen was located by traversing the slide at 1 mm spacing until a minimum of 200 grains were identified. Identification was aided by the use of pollen identification keys (Faegri et al., 1989, Moore et al., 1991), pollen picture atlases (Reille, 1992, Reille, 1995, Reille, 1998) and referral to the pollen reference collection held within the Long-Term Ecology Lab. The pollen sum (percentage) is based on a minimum of 200 terrestrial pollen grains (including Cyperaceae) that were counted for each level. Aquatic pollen percentages are based on the sum of the terrestrial pollen and aquatic pollen.

In the papers presented in this thesis the pollen record is split into two sections as follows: **Chapter 5** covers the section of the sedimentary sequence from 600- 544 cm (~ 112 to ~ 104 ka BP) and **Chapter 6** (544-338 cm, ~ 104 to ~ 35 ka BP). The entire pollen sequence is presented here in Fig. 3.3 for reference. The very low concentrations and degradation of the pollen grains from 700-601 cm (~ 123 to ~ 112 ka BP) prevented any extraction of material from this section of the sedimentary sequence. The lower section of the sediment core (700-601 cm) and the levels 600-580 cm (~ 112 to ~ 109 ka BP) where pollen concentrations as described in **Chapter 5** are very low is comprised of calcium carbonate ($CaCO_3$); it is probable that the highly alkaline sediments resulted in the deterioration of the pollen. The variations of the dominant pollen types as detailed in **Chapters 5** and **6** indicate that the composition of the terrestrial vegetation community oscillated between an open forest/grassland and steppe during the last glacial period. Additionally, changes in aquatic and semi-aquatic herbaceous taxa suggest significant fluctuations in the habitat structure in proximity to the core site within the wetland.

3.4 Core contamination

The discovery of an oxidised steel nail in the sediment core at 337 cm during sub-sampling for fossil-pollen indicates either extensive physical reworking of the sediment above this depth or contamination during the coring procedure. If this is an artefact of reworking of sediment post initial burial it may have occurred due to one of the following processes. Firstly, due to the physically dynamic nature of the Aammiq Wetland sediment basin that at some time in the past had the Nahr el Hafir River (Fig. 2.2) flowing through it (Walley, 1996) cutting channels and re-depositing sediment. Secondly, given that today almost the entire Bekaa Valley is given over to agricultural production (FAO, 2008), it is

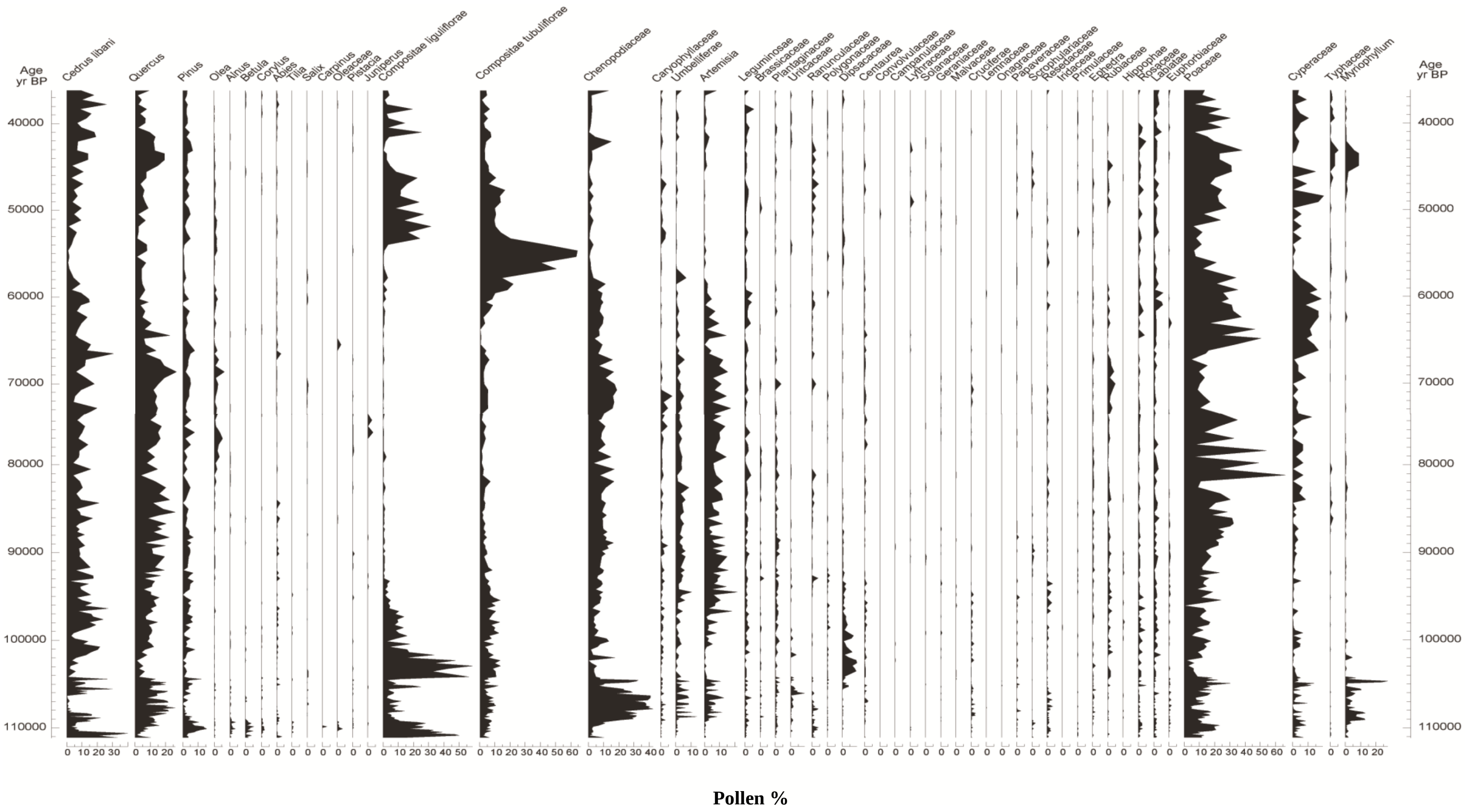

Figure 3.3 Pollen percentage diagram for Aammiq Wetland constructed using PSIMPOLL version 4.27 (Bennett, 2009) for the entire section of the core (600-338 cm) analysed in this thesis plotted against age. The bottom of the pollen sequence dated at ~ 112 ka BP corresponds to the end of the last interglacial period and the top the later stages of the last glacial period (~ 35 ka BP).

possible that areas of the current wetland were in the past also cultivated and then left fallow and allowed to re-flood, a process that has occurred in the recent past to expand the area of the wetland (see Fig. 2.2) (Storey, 2007). However, if the sediment has been extensively re-worked it would be highly likely that the AMS and OSL dates obtained from the sediment below 337 cm would be subject to age-reversal and therefore the age-depth model they are used to construct would not display the linearity that is observed. However the steel nail became deposited into the sediment its presence and therefore the potential for contamination precluded analysis of the sediment record above 337 cm. In conjunction with the low concentrations of fossil pollen below 600 cm highlighted above analysis of the sediment core detailed in this thesis is restricted to 600-338 cm of the 7 m sediment core extracted from Aammiq Wetland.

3.5 Stable isotopes of carbon and total organic carbon content

Analysis of changes in $\delta^{13}C$ of the sediment (**Chapter 5**) was undertaken to demonstrate changes in plant type based on the photosynthetic pathway (C_3 or C_4) (Fontugne and Calvert, 1992). As C_4 plants predominately persist within warm to hot climatic zones (Ehleringer et al., 1991) the carbon isotope signal also potentially has the ability to act as an indicator of climatic change. Changes in total organic carbon (TOC) was also determined to indicate the level of plant productivity within the lake and its surrounding terrestrial landscape (Meyers and Ishiwatari, 1993, Meyers, 1994). Analysis of the sediments was conducted at the Godwin Laboratory for Palaeoclimate Research, Cambridge. Prior to analysis sediments were decarbonized with HCl 7%, homogenised and then dried. Analysis for TOC and isotopic composition of carbon ($\delta^{13}C$) was conducted using a Costech Elemental Analyser attached to a Thermo MAT 253 mass spectrometer in continuous flow mode.

3.6 Geochemistry

Electron microscopy combined with EDXS (Webb et al., 2000) was undertaken to analyse the elemental composition of lake sediments (Banfield et al., 1991, Fortin et al., 1993, Hupfer et al., 1995). Composition of the major and trace elements of the sedimentary sequence were determined at 2 cm intervals (600-338 cm) using energy dispersive x-ray spectroscopy (EDXS) at the University of Oxford on a Hitachi TM-3000 electron microscope (EM) fitted with an internal energy dispersive spectrometer (EDXS). The results of the geochemistry are shown in Fig. 3.4 and discussed in **Chapter 4.**

3.7 Loss-on-ignition

Loss-on-ignition (LOI) conducted on 600-338 cm of the sediment core (detailed in **Chapter 4)** was used to estimate both the organic and carbonate content of lake sediments (Bengtsson and Enell, 1986). Bulk sediments (1 cm^3) were extracted at 2 cm intervals and oven-dried in a Carbolite® chamber furnace (model ELF11/14B) at 105 ºC for 24 hours, to remove organic matter at 550 ºC for 4 h and carbonate at 950 ºC for 2 h (Heiri et al., 2001). Loss-on-ignition of organic matter ($LOI_{550\,^{\circ}C}$) as a percentage of the dry weight ($DW_{105\,^{\circ}C}$) of the sediment in grams was then calculated using the equation:

$$LOI_{550\,^{\circ}C} = ((DW_{105\,^{\circ}C} - DW_{550\,^{\circ}C})/DW_{105\,^{\circ}C})*100$$

Loss-on-ignition of carbonate was calculated using the equation:

$$LOI_{950\,^{\circ}C} = ((DW_{550\,^{\circ}C} - DW_{950\,^{\circ}C})/DW_{105\,^{\circ}C})*100$$

3.8 Magnetic susceptibility

Magnetic susceptibility (MS) is a geochemical proxy that provides a rapid and non-destructive method (Peck et al., 1994, Lecoanet et al., 1999) to measure the magnetic minerals in an environmental material (Dearing, 1999). Variability in MS values in lake sediments can be indicative of the different erosional processes and rates within a lake catchment (Thompson et al., 1975, Oldfield et al., 1978). Magnetic susceptibility (MS)

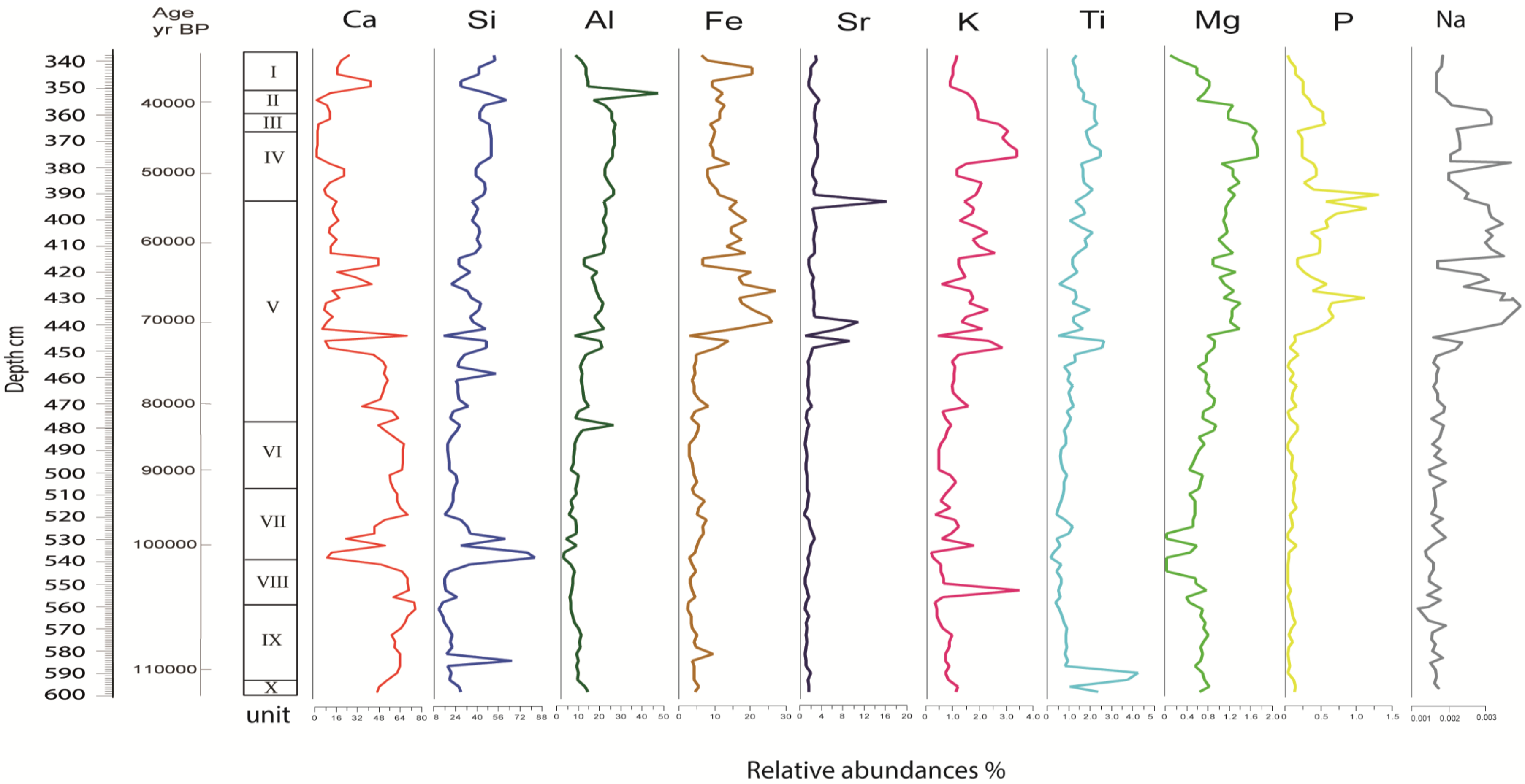

Figure 3.4 Energy dispersive x-ray spectroscopy (EDXS) results from the Aammiq Wetland sediment core indicating changes in the relative abundances (percentages) of selected elements (Ca, Si, Al, Fe, Sr, K, Ti, Mg, P, Na) for the entire section of the core (600-338 cm) analysed in this thesis. The element abundances are plotted against age and depth and differentiated into units to aid presentation of the results detailed in **Chapter 4**.

measurements were determined using a Bartington Instruments MS2 meter fitted with a MS2C core scanning sensor. The sediment core (all 700 cm) was scanned at 1 cm intervals the results and analysis of the MS for the section of core 600-338 cm is detailed in **Chapter 4**. To account for magnetic drift the mean of three measurements were taken.

3.9 Macro and microcharcoal

Analysis of charcoal from 544-338 cm (Fig. 3.5) as described in **Chapter 7** was undertaken to enable the reconstruction of long-term variations in the occurrence of fire and the effects of fire on ecosystems (Whitlock and Larsen, 2001). The presence of charcoal particles has been determined according to (Swain, 1973) as those particles being angular, opaque and uniformly black. The number of microcharcoal particles < 150 µm was determined by a count of its occurrence on pollen slides, this is a commonly employed method (Finsinger and Tinner, 2005). Charcoal fragments and *Lycopodium* were counted until the combined ∑=200 (Tinner and Hu, 2003). The macrocharcoal > 150 µm fraction of the sediment was extracted during the sieving stage of pollen preparation. The counting of the macrocharcoal was then conducted in an uncovered petri dish under a stereo light microscope.

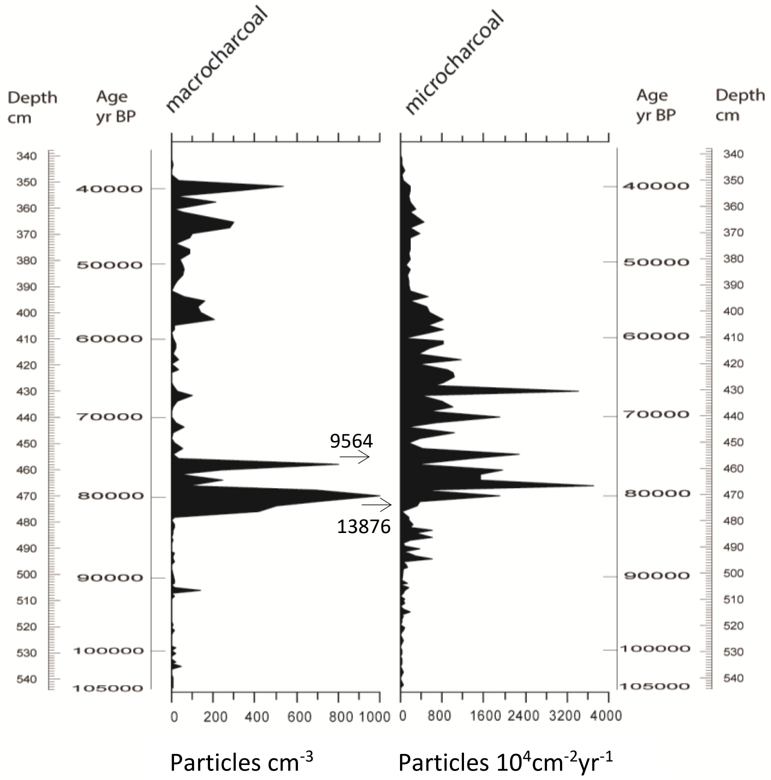

Figure 3.5 Macrocharcoal concentration and microcharcoal accumulation data for Aammiq Wetland plotted against age and depth.

3.10 Statistical analysis

3.10.1 Principal component analysis

Principal component analysis (PCA) is a statistical method used on multivariate data (Jolliffe, 2002). Applied to large ecological data sets of interrelated variables PCA has the ability to summarise the primary patterns of variation (Davis, 1986) by reducing the number of variables in a data set into two or three principal components (Webb et al., 1978). This approach enables the maximum variability between data points to be identified (Davis, 1986) and displayed visually (Tipping and Bishop, 1999). Principal component analysis was applied to the pollen data using the software C2 version 1.7.2 (Juggins, 2011).

3.10.2 Cross-correlation

Cross-correlation analysis of the pollen and charcoal time series data was used to measure the correspondence between variables over time (Green, 1981). Cross-correlation was applied to the charcoal and pollen time series data to determine if there was any correlation between peaks in charcoal and changes in individual plant taxa and/or plant assemblages and to determine the extent of any lead or lags. Cross-correlation was conducted using the software PAlaeontological STatistics (PAST) version 2.17 (Hammer, et al., 2001).

3.10.3 Spectral analysis

Spectral analysis is a statistical method used to determine the presence of periodic characteristics in time series data (Pisias et al., 1973). Conducted on the charcoal data spectral analysis was used to ascertain if there was any periodicity in peak charcoal influx events. Spectral analysis was conducted using the software PAST version 2.17 (Hammer et al., 2001).

3.10.4 Rarefaction analysis

The number of taxa in any pollen sample is dependent on the number of pollen grains counted and this varies between samples (Birks and Line, 1992). Rarefaction analysis compensates for differences in the number of pollen grains counted between samples by predicting what the number of pollen taxa would be if the pollen count for each sample was the same (Seppä and Bennett, 2003). Rarefaction analysis has the potential to provide insights into changes in taxonomic diversity and its potential causes (Seppä and Bennett, 2003). Rarefaction analysis was applied to the pollen time series data using PSIMPOLL version 4.27 (Bennett, 2009).

3.10.5 Rate-of-change analysis

Rate-of-change analysis (Grimm and Jacobson, 1992) was performed on the pollen data so that the rate of vegetation change over time could be estimated. The rate-of-change analysis that was applied to the pollen data using PSIMPOLL version 4.27 (Bennett, 2009), measures the difference between adjacent samples using chord distance and then divides the value obtained by the time lapse between the samples (Bennett and Humphry, 1995); the greater the value obtained the larger the rate-of change.

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Chapter 4: Landscape erosion, karstic activity and the development of a wetland in the southern Bekaa Valley, Lebanon during the last glacial period

4.1 Abstract

We present the results of the first multi-proxy sedimentological analysis of a sedimentary sequence recovered from Aammiq Wetland in the Bekaa Valley, Lebanon that spans the very end of the last interglacial and the last glacial period (~ 112 to ~ 35 ka BP). The high carbonate content of the sediment that accumulated between ~ 112 to ~ 85 ka BP was due to the significant influx of karstic spring water into the wetland. This occurrence suggests that three components of the hydrological system were present during this time: firstly, there was intense water circulation through the karstic aquifer; secondly significant precipitation over the Lebanon Mountains; and thirdly the presence of a shallow open water lake in the Bekaa Valley. This was a time of high forest cover and stable soils as indicated by high arboreal pollen and low concentrations of siliclastic elements. This period of catchment stability was interrupted by a protracted period of high erosion (~ 108 to ~ 105 ka BP) before returning to the stable catchment state. A further shift in the hydrological system around ~ 84 ka BP when wet climatic conditions were replaced by low water circulation (through the karstic system) and a drop in lake levels fostered the development of a wetland/swamp. Around ~ 60 ka BP during the full glacial, there was another period of high erosion concurrent with a decline in precipitation and terrestrial vegetation cover. From ~ 39 ka BP as precipitation over the Lebanon Mountains increased water circulation through the karstic aquifer recommenced and carbonates were once again deposited in the wetland.

Keywords: *Bekaa Valley*, *last glacial period*, *sediments*, *erosion*, *karstic system, lake, wetland/swamp*

4.2 Introduction

The topography of Lebanon's interior is comprised of the NNE-SSW aligned Lebanon and Anti-Lebanon Mountain ranges that are separated by the elevated upland basin of the Bekaa Valley (Walley, 1996, Nader and Swennen, 2004). The bedrock of these mountains is predominantly comprised of Jurassic limestone with some exposed beds of Cretaceous sandstones (Edgell, 1997, Walley, 1998). The soils that have developed on these calcareous rocks are typically shallow lithic leptsols (Asmar, 2011) on the higher slopes and deeper inceptisols (Vogiatzakis, 2012) and terric anthrosols (Asmar, 2011) on the lower slopes. The combination of the steep sloping terrain, calcareous geology and the relatively shallow soils, ensures that erosion is a very active process on the mountains of Lebanon (Vogiatzakis, 2012). This is a process that has been exacerbated in recent times by poor agricultural practices (Bou Kheir et al., 2006, Bou Kheir et al., 2008) and the logging of timber (Choueiter and Ucenic, 2007).

Erosion of the mountainous terrain is closely coupled to the significant levels of rainfall and snow that falls on the mountain slopes (Khawlie et al., 2002). This high level of precipitation in combination with the predominantly calcareous geology has caused the growth of well-developed karstified carbonate strata (Edgell, 1997), that includes a number of aquifers (Edgell, 1997) and associated surface springs (Walley, 1996, Edgell, 1997). The infiltration of water into karstic fractures and conduits that replenish subsurface aquifers causes much of the surface rainfall or snow melt to be lost quickly (Khawlie et al., 2002). Karst therefore plays a fundamental role in the hydrological cycle of the Lebanon Mountains and the Bekaa Valley (Edgell, 1997). Today in the Bekaa Valley karstic spring discharge and the extraction of groundwater provide a majority of the water used for the cultivation of crops (Karam et al., 2007, Asmar, 2011) and as such the well-being of people living in the Valley is closely tied to the availability of water in the system.

The combined effects of water abstraction and declining precipitation during the late-Holocene (Walley, 1996) have progressively reduced the extent of the wetlands in the southern Bekaa Valley (Walley, 1996, Storey, 2007). As a consequence, the wetland that persists today is ephemeral (Walley, 1996, Storey, 2007). It is generally only full of water from January to March, with water levels declining from May; in a typical year, the wetland is totally dry from late July until December (Storey, 2007). The area that is continuously flooded, fed by the influx of spring water during the wet season, is generally limited to approximately 70 ha (Storey, 2007), although this can temporarily expand in area up to 300 ha during periods of heavy rainfall and snow melt.

Whilst our knowledge about the historical dynamics of karst hydrology, soils and associated landscape processes in this watershed is relatively well understood, little is known about how these processes varied over glacial to interglacial time scales and how in turn this affected the wetlands that occupied the southern Bekaa Valley.

This study presents the results of a multi-proxy sedimentological study (geochemistry, loss-on-ignition and sediment type and magnetic properties) conducted on a new sedimentary sequence from a small wetland (Aammiq) in the Bekaa Valley, Lebanon that was deposited at the very end of the last interglacial and during the last glacial period (~ 112 to ~ 35 ka BP). The aims of this paper are: (1) to investigate changes in chemical, magnetic and physical properties of the sediment over time (2) to infer what were the potential climatic and environmental processes driving these sedimentary changes, and (3) to determine how these changes in sedimentary properties and climate affected the wetland in the southern Bekaa Valley.

4.3 Modern setting

Aammiq Wetland is situated on the western side of the southern Bekaa Valley, Lebanon (33°43'50''N, 35°47'10''E) at 860 meters above sea level (m a.s.l) (Fig. 4.1). Today the wetland is ephemeral (Walley, 1996) expanding during late-winter and spring to typically encompass ~ 70 ha (Storey, 2007). It is a small remnant of far more extensive wetland that existed until the 1940's (Walley, 1996), that was progressively drained during the 1940-50's and converted to agricultural production (Walley, 1996, Storey, 2007). Today the flat fertile valley plain is almost entirely given over to the cultivation of crops.

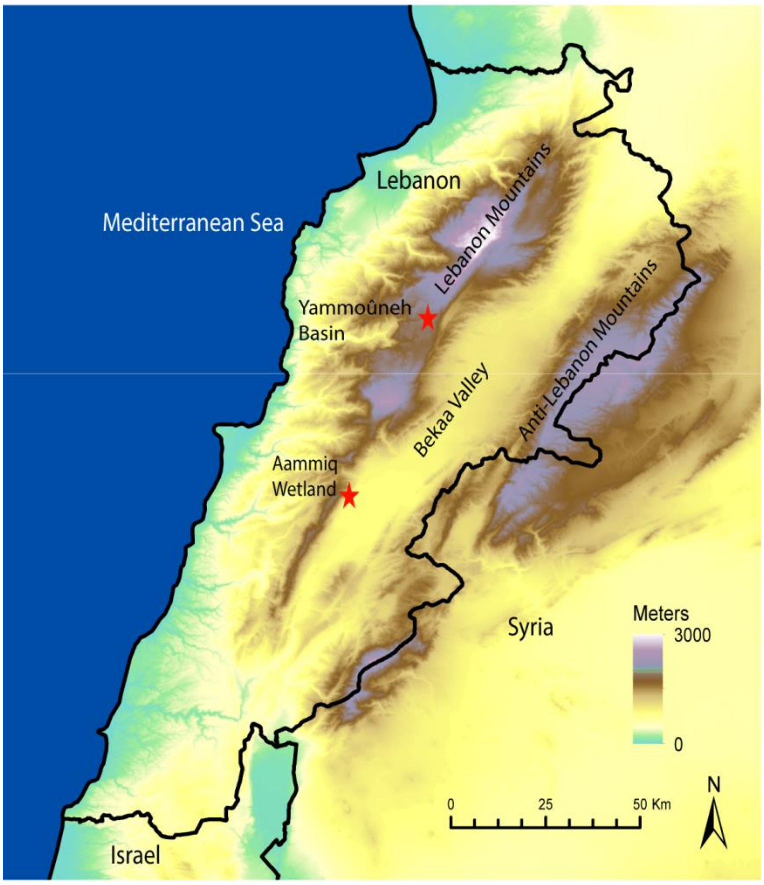

Figure 4.1 Topographic map of Lebanon showing location of Aammiq Wetland (this study) and Yammoûneh Basin the location of previous palaeoenvironmental studies in Lebanon conducted by Develle et al. (2011) and Gasse et el. (2011) that cover the last glacial period.

Geology and geomorphology

The major geological structures of this region are related to the NNE-SSW aligned Yammoûneh Fault System (Walley, 1996, Walley, 1998), which forms part of the Dead Sea Transform Fault that extends from the NW tip of the Red Sea to the Taurus Mountains of

Anatolia. In the Aammiq area the Yammoûneh fault is comprised of at least three parallel branches (Walley, 1996), see figure. 4.2. The Lebanon Mountains (max. elevation Qurnat as Sawda 3088 m) that lay approximately parallel to this fault system, are the result of plate deformation (Dubertret et al., 1970). The bedrock of the these mountains is predominately comprised of Cretaceous and Jurassic limestones (Walley, 1996, Edgell, 1997), with some deformed basal Cretaceous sandstone located on the lower slopes of the Jebel Barouk-Jebel Niha Massif (Edgell, 1997), the flank of which end at the wetlands edge (Walley, 1996). The shallow soils that have developed on the mountain slopes are predominately composed of lithic leptosols (Asmar, 2011) and inceptisols (Vogiatzakis, 2012), with terric anthrosols (Asmar, 2011) having developed in areas that have been subjected to significant human activity. On the lower slopes of Jebel Barouk is an alluvial fan > 1.5 km in width that is comprised of a mixture of sandstone and limestone clasts (Walley, 1998). In the Bekaa

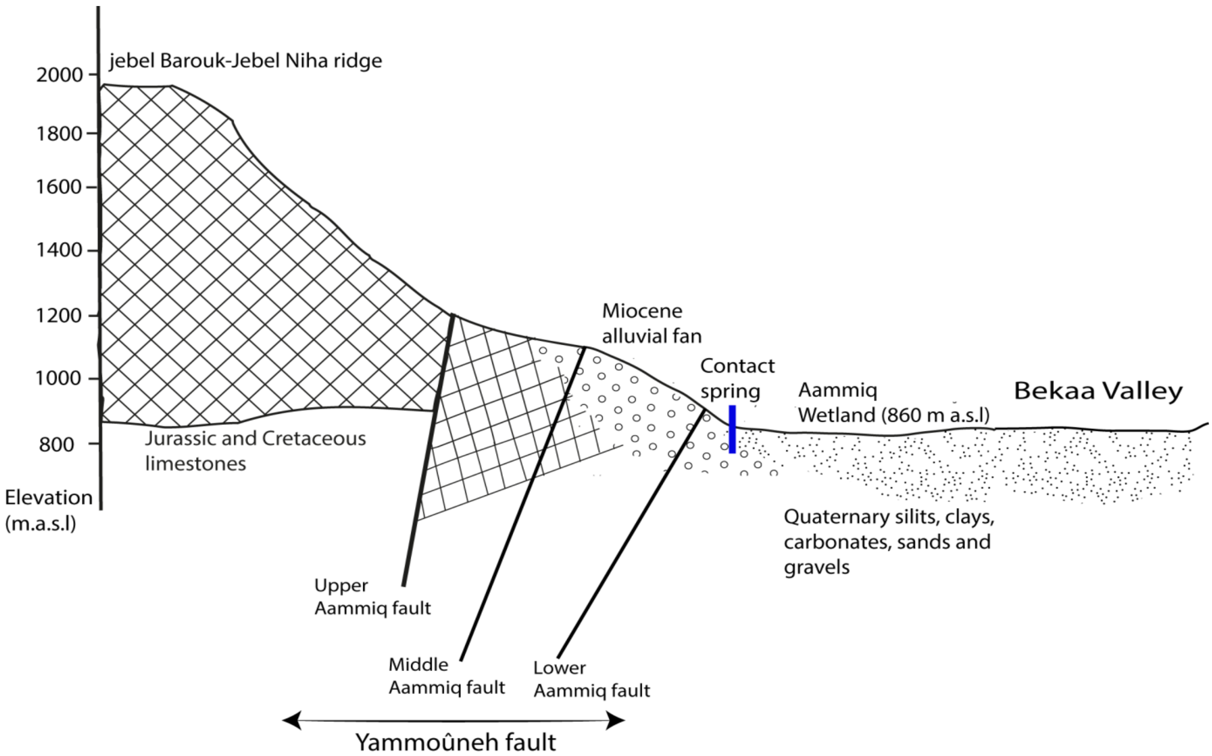

Figure 4.2 Geological cross-section of the Lebanon Mountains in the Aammiq region, modified from Walley (1998). Also shown is the position of the Aammiq Contact Spring (Nabaa Aammiq) the main source of water input into the wetland.

Valley itself the sediments are comprised mainly of alluvial deposits consisting of silts, clays, sand and gravel that were laid down during the Quaternary period (Walley, 1998), and are probably in excess of 1 km in depth in parts of the wetland (Walley, 1998).

Climate

The climate is Mediterranean in type with warm dry summers and cool wet winters. Average temperatures in the southern Bekaa Valley range from 2 °C in winter to 34 °C in summer (Ministry of Environment/Med Wet Coast Project, 2004). Average annual precipitation (1954-2002 long term average, as recorded at Tal Amara meteorological station, 30 km north of Aammiq) is 600 mm yr^{-1} (Karam et al., 2007). The low level of precipitation in the Bekaa in comparison to Lebanon's coastal areas, which receive an average of between 700-1000 mm yr^{-1} (Asmar, 2011) is due to the rainshadow effect of the Lebanon Mountains which reduces the moisture content of airflows that pass from the Mediterranean Sea over to the Bekaa in the east (Sene et al., 1999). Precipitation typically occurs on 80-90 days a year, mainly between November and April (Storey, 2007) due to the passage of Cyprus Lows (extratropical cyclones) that are the main source of precipitation into the eastern Mediterranean (Alpert et al., 1990, Shay-el and Alpert, 1991). On the higher elevations of the Lebanon Mountains average annual precipitation can exceed 1600 mm yr^{-1} (Asmar, 2011), with a majority of the precipitation falling as snow (FAO, 2008). On the mountain slopes average annual temperature of the coldest months (December, January, February) is below 0 °C and the warmest month (July) exceed 30 °C (Gasse et al., 2011).

Hydrological setting

Water input into Aammiq Wetland occurs through a combination of direct precipitation and the inflow of spring water (Walley, 1996); with the greatest input of water being derived from numerous contact springs in the area (e.g Nabaa Barouk and Nabaa Aammiq) that are fed from a karstified aquifer situated within the slopes of the Jebel Barouk-

Jebel Niha Massive (Walley, 1996, Edgell, 1997, Storey, 2007). Water levels in the subterranean karstic network are recharged from a combination of snowmelt and rainfall (Walley, 1996, Edgell, 1997). Due to the high transmissivity of the aquifer, water residence time is limited to one or two seasons and thus spring discharge declines rapidly during the dry summer and autumn months (Edgell, 1997).

4.4 Materials and methods

A 7 m sediment core was extracted from Aammiq Wetland (33°43'51.47"N, 35°47'08.80"E) using a Russia corer. The core was wrapped in cellophane and aluminium foil to retain moisture and then encased in black pvc tubing to minimise the potential for light to penetrate into the sediments so that optically stimulated luminescence dating could be conducted.

Age control

Chronology of the core was determined by a combination of radiocarbon dating conducted at ^{14}CHRONO Radiocarbon Laboratory, Queen's University Belfast, Belfast, UK and optically stimulated luminescence dating (OSL) performed at the Oxford Luminescence Laboratory at the University of Oxford, Oxford, UK.

As the greatest input of water into Aammiq Wetland is from springs fed from kartisfied aquifers (Walley, 1996) the radiocarbon dates obtained from bulk sediments are potentially subject to the 'reservoir effect' of geologically old carbon entering the aquatic system. This produces age estimates that are too old by up to several thousand years (Meadows, 2005). An attempt to circumnavigate this effect was made by extracting the terrestrial fossil pollen fraction from the sediment for radiocarbon dating following the protocol of Brown et al. (1989). Unfortunately the concentrations of pollen sent to ^{14}CHRONO and hence carbon levels were too low to produce any signal. Therefore with an understanding that any dates obtained by atomic mass spectrometry (AMS) will potentially have a significant error, dating was conducted on bulk sediments. In the laboratory, 0.5 g of bulk sediment was extruded from two depths: 258 and 360 cm and sent to ^{14}CHRONO for sample preparation and Accelerator Mass Spectrometric analysis (AMS).

For OSL dating in the laboratory under yellow light conditions, 1 cm^3 of sediment was extracted from 30 levels. The sediment was then treated with HCl (30%) to remove carbonates and hydrogen peroxide H_2O_2 (35-50%) to remove organics. The remaining sediment was then sieved at 150 μm to separate the fine and coarse fractions. The coarse fraction was then treated with HF (58-62%) to remove silicates and HCl to remove colloidal silica and silicofluorides created by the HF reaction. The coarse quartz fraction that remained from only 4 levels (690, 544, 278 and 219 cm) was then washed in H_2O and placed in a drying oven for 24 hr to allow evaporation of the remaining water. The fine grain (5-15 μm) fraction from 486 cm was then wet sieved and subjected to density separation through flotation in H_2O before being placed in a drying oven for 24 hr. The dried quartz fractions were analysed using the Single Aliquot Regenerative (SAR) protocol of Murray and Wintle (2000 and 2003). Measurements were made on a Risø TL-DA-15 reader using blue diode illumination and u-340 filters at the Oxford Luminescence Laboratory (OLD). A further 1 cm^3 of sediment was extracted from the core at the same levels to be sent for analysis by ICP-MS to estimate the natural dose rate.

Calibration of the bulk sediment radiocarbon dates (^{14}C ages) to years before present (cal. yr BP) was performed using CLAM's default northern hemisphere (atmospheric) calibration curve that uses IntCal09 (Reimer et al., 2009). The Age-depth model for the sedimentary sequence combining the calibrated radiocarbon dates and the OSL dates was then constructed using linear interpolation in the software program CLAM version 2.1 (Blauuw, 2010) run in R programming language (2010).

Elemental composition

The relative concentrations of the major elements were measured at 2 cm intervals by scanning electron microscope-energy dispersive spectroscopy (SEM-EDS) at the University of Oxford, Material Characterisation Services, Oxford, UK Prior to measurement the sediments were dried at 105 °C for 24 hr, homogenised using a mortar and pestle. The sediment was then compressed into a pellet, adhered to a carbon tab and mounted on an aluminium stub. Measurements were performed using a Hitachi TM 3000 with an integrated Bruker Quantax 70 EDS. An electron beam was used to excite X-ray emissions that are detected by a peltier cooled silicon drift detector (SDD). The analytical settings were set at 15 kv in order to detect Si, Ca, Al, Fe, Sr, K, Ti, Mg, P and Na.

Sedimentary magnetism

Magnetic susceptibility (MS) was used to measure the magnetic minerals in the bulk sediments to infer erosional processes in the catchment (Thompson et al., 1975, Dearing, 1999). The low field magnetic susceptibility $\kappa_{LF,}$ was measured using a Bartington Instruments MS2 meter fitted with a MS2C core scanning sensor. The sediment core was scanned at 1 cm intervals resulting in a total of 262 readings. To account for magnetic drift the mean of three measurements was taken from each level.

Loss-on-ignition

Loss-on-ignition (LOI) was used to estimate both the organic and carbonate content of lake sediments (Bengtsson and Enell, 1986). Bulk sediment (1 cm^3) was extracted at two centimeter intervals resulting in a total of 130 samples. Samples were oven-dried in a Carbolite® chamber furnace (model ELF11/14B) at 105 ºC for 24 hours, to remove organic matter at 550 ºC for 4 hours and carbonate at 950 ºC for 2 hours (Heiri et al., 2001). Loss-on-ignition of organic matter ($LOI_{550\ °C}$) as a percentage of the dry weight ($DW_{105\ °C}$) of the sediment in grams was then calculated using the equation:

$$LOI_{550\,°C} = ((DW_{105\,°C} - DW_{550\,°C})/DW_{105\,°C})*100$$

Loss-on-ignition of carbonate was calculated using the equation:

$$LOI_{950\,°C} = ((DW_{550\,°C} - DW_{950\,°C})/DW_{105\,°C})*100$$

4.5 Results

Chronology

Results of the radiocarbon and OSL dating are presented in table 4.1 and 4.2 respectively and Fig. 4.3. The section of the 7 m sedimentary sequence recovered from Aammiq Wetland covered by this study (600-338 cm) spans the very end of the last interglacial and the last glacial period (~ 112 to ~ 35 ka BP).

Sample depth (cm)	Laboratory number	Sample Material	^{14}C age yr. BP	2σ Calibrated age (yr BP)
258	UBA-20316	Bulk sediment (Clay/organic rich)	12836 ± 69	14939-15882
360	UBA-20315	Bulk sediment (Clay/organic rich)	37205 ± 745	41058-43108

Table 4.1 Radiocarbon dating results from the Aammiq Wetland sedimentary sequence. ^{14}C ages were calibrated using IntCal09 (Reimer et al., 2009).

Sample depth (cm)	Lab code	Grains accepted to calculate D_e	Minimum Maximum grain size (µm)	K (%)	U (ppm)	Th (ppm)	Cosmic dose rate (Gy/ka)	Total dose Rate (Gy/ka)	D_e (Gy)	Age (ka)
690	OxL 2270	16	75-125	1.04 ± 0.05	4.06 ± 0.46	15.69 ± 1.56	0.10 ± 0.01	2.87 ± 0.18	355.56 ± 30.03	124 ± 13.2
544	OxL 2268	18	75-125	0.27 ± 0.007	1.91 ± 0.08	3.92 ± 0.25	0.12 ± 0.01	1.01 ± 0.05	105.6 ± 8.25	104.6 ± 9.5
486	OxL2267	12 *	5-15	0.38 ± 0.02	3.4 ± 0.2	5.2 ± 0.26	0.10 ± 0.01	1.86 ± 0.17	163.0 ± 5.57	87.8 ± 8.6
278	OxL 2271	16	75-125	0.25 ± 0.12	3.4 ± 0.34	1.0 ± 0.1	0.16± 0.01	0.81 ± 0.05	16.28 ± 1.02	20.1 ± 1.7
219	OxL 2263	24	75-125	0.48 ± 0.02	4.52 ± 0.45	10.32 ± 1.0	0.18 ± 0.01	2.21 ± 0.15	11.38 ± 0.48	5.1± 0.4

Table 4.2 Results of optically stimulated luminescence dating. * D_e (equivalent dose) was calculated from 12 aliquots of fine grain quartz, assuming a moisture content of 15 ± 5%. Moisture content of all other samples was estimated to be 10 ± 3%.

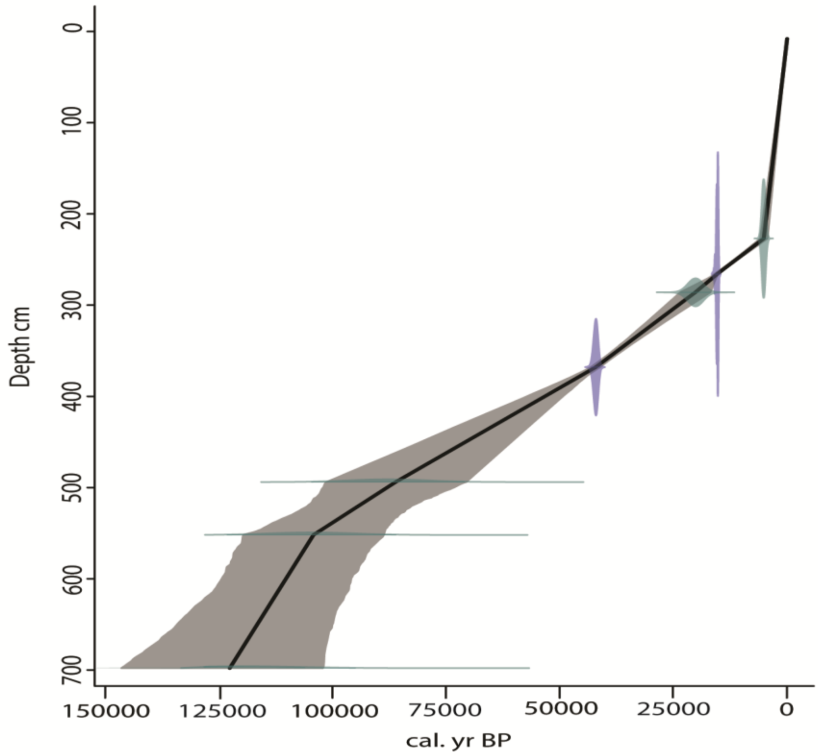

Figure 4.3 Age-depth graph undertaken in CLAM version 2.1 (Blaauw, 2010) using linear interpolation. Black line represents the ‘Best fit’ model and the grey shaded area 95% confidence limits. The probability distribution of the calibrated radiocarbon dates are shown as blue reduced plots, and the probability distribution of the OSL dates as green reduced plots.

Stratigraphy of the sediments

The sedimentary sequence can be divided into 10 major lithological units according to changes in sediment type. A description of these sedimentary units is given below and also presented in Fig. 4.4. Approximate ages assigned to these units are extrapolated from the age-depth model. Units II (360-352 cm, ~ 42 to ~ 39 ka BP), IV (396-377 cm, ~ 53 to ~ 44 ka BP), VI (508-480 cm, ~ 93 to ~ 85 ka BP) and X (600-595 cm, ~ 112 to ~ 111 ka BP) consist of dark grey to olive grey silty clays. Units VII (540-509 cm, ~ 103 to ~ 94 ka BP) and IX (594-571 cm, ~ 111 to ~ 109 ka BP) are predominately comprised of powdery light grey lacustrine carbonates that contain varying amounts of plant debris. Units I (351-338 cm, ~ 39 to ~ 35 ka BP) and VIII (571-540 cm, ~ 108 to ~ 104 ka BP) consist of sand and contain significant quantities (≤ ~ 50% volume) of coarse (≤ Ø15 mm) limestone gravels. At the end of unit I and the start of unit VIII the sediments also contain calcareous silt, that increases in abundance through unit I and declines in unit VIII. Unit V (479-396 cm, ~ 84 to ~ 54 ka BP) is a dark brown to almost black organic rich clay rich in plant material. Unit III (376-361 cm, ~ 44 to ~ 42 ka BP) is a black band of peat.

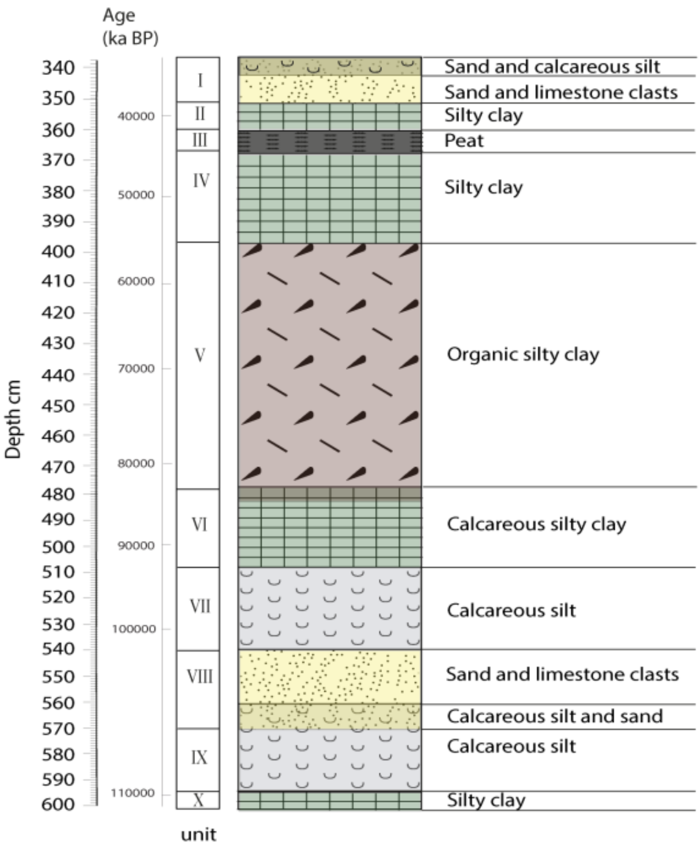

Figure 4.4 Stratigraphic profile of the lithology of the Aammiq Wetland sediment core plotted against depth and age and differentiated into units to aid presentation of the results detailed above. The sediment is comprised of four primary components: calcareous silt, sand, clay and organic matter.

Geochemical analysis

The results of the elemental analysis are illustrated in Fig. 4.5. Calcium is the predominant element in units II, VI, VII, IX and X. Concentrations of Si reach their maximum levels during unit VIII that is associated with the sedimentation of quartz. The abundances of Si, Al, K and Ti show a gradual increase during unit VII that corresponds to the decrease in Ca. High concentrations of Fe and P occur within the organic rich sediment unit V. An increase in the concentrations of Si, Al, K, Ti, Mg and Na occurs during units IV and V. In unit I Si and Ca increase and Fe decreases.

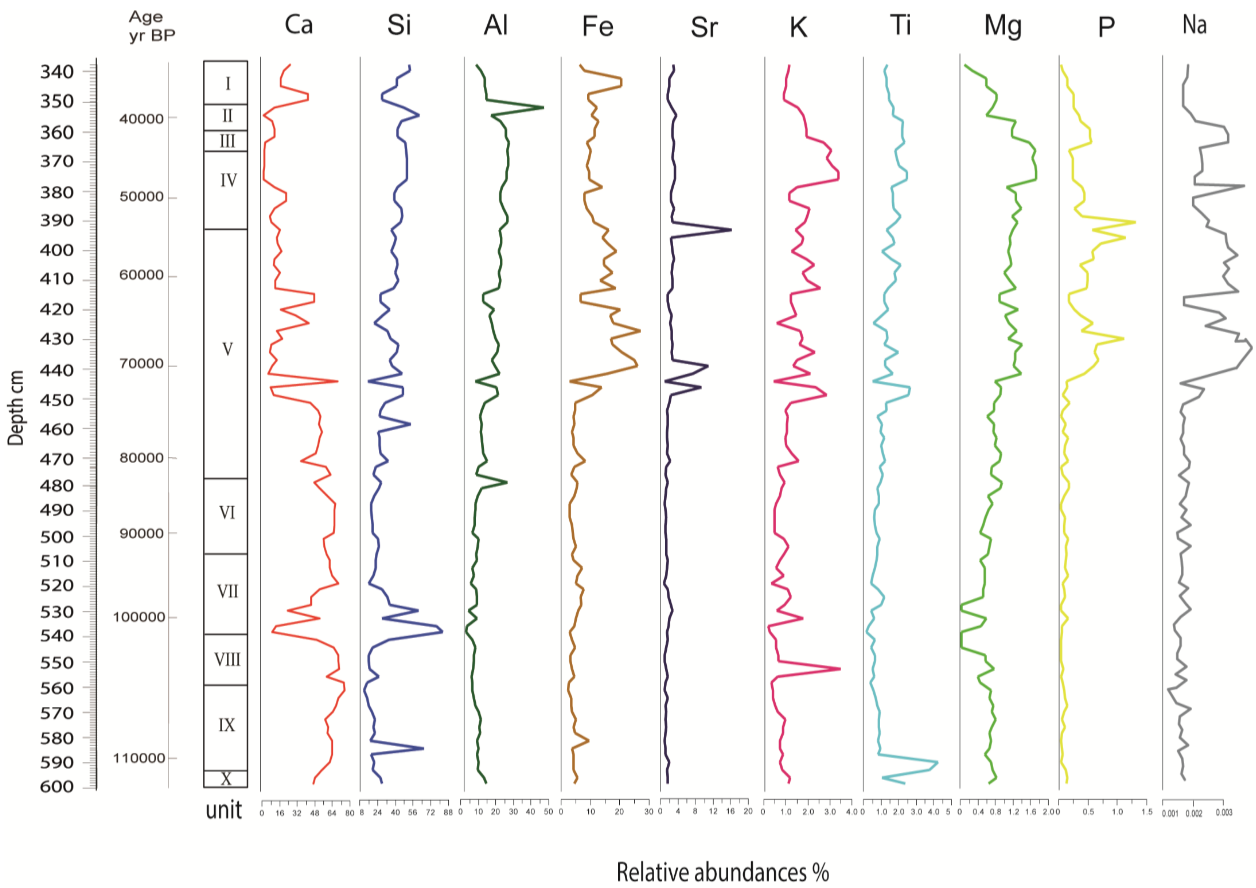

Figure 4.5 Energy dispersive x-ray spectroscopy (EDXS) results from the Aammiq Wetland sediment core indicating changes in the relative abundances (percentages) of selected elements (Ca, Si, Al, Fe, Sr, K, Ti, Mg, P, Na). The elements are plotted against depth and age and differentiated into units to aid presentation of the results detailed above. Note the decline in Ca and concomitant increase in all other elements during unit V at ~ 74 ka BP.

Magnetic susceptibility and loss-on-ignition

The results of the MS and LOI analyses are presented in Fig. 4.6. All MS measurements are displayed in SI $\times 10^{-5}$ units. The magnetic drift between the three measurements conducted at each level was not significant (0 and 1.8). The MS values are high during unit X (range of 14.5-9 SI units); MS values then decline to < 1 by the start of unit VIII. Magnetic susceptibility then increased rapidly to 16.5 during unit VII and remained high (range of 16.5-9.7) until the start of unit VI. The MS values then declined abruptly to 3.4 and then immediately started to increase (≤ 7.9). During unit V MS values declined to 0 and remained low (< 6) with the exception of a spike (14.6) during the later period of this unit. Magnetic susceptibility values during unit II increased abruptly (12) before declining (< 4) and then increasing again during unit I attaining the highest values of the entire record (20).

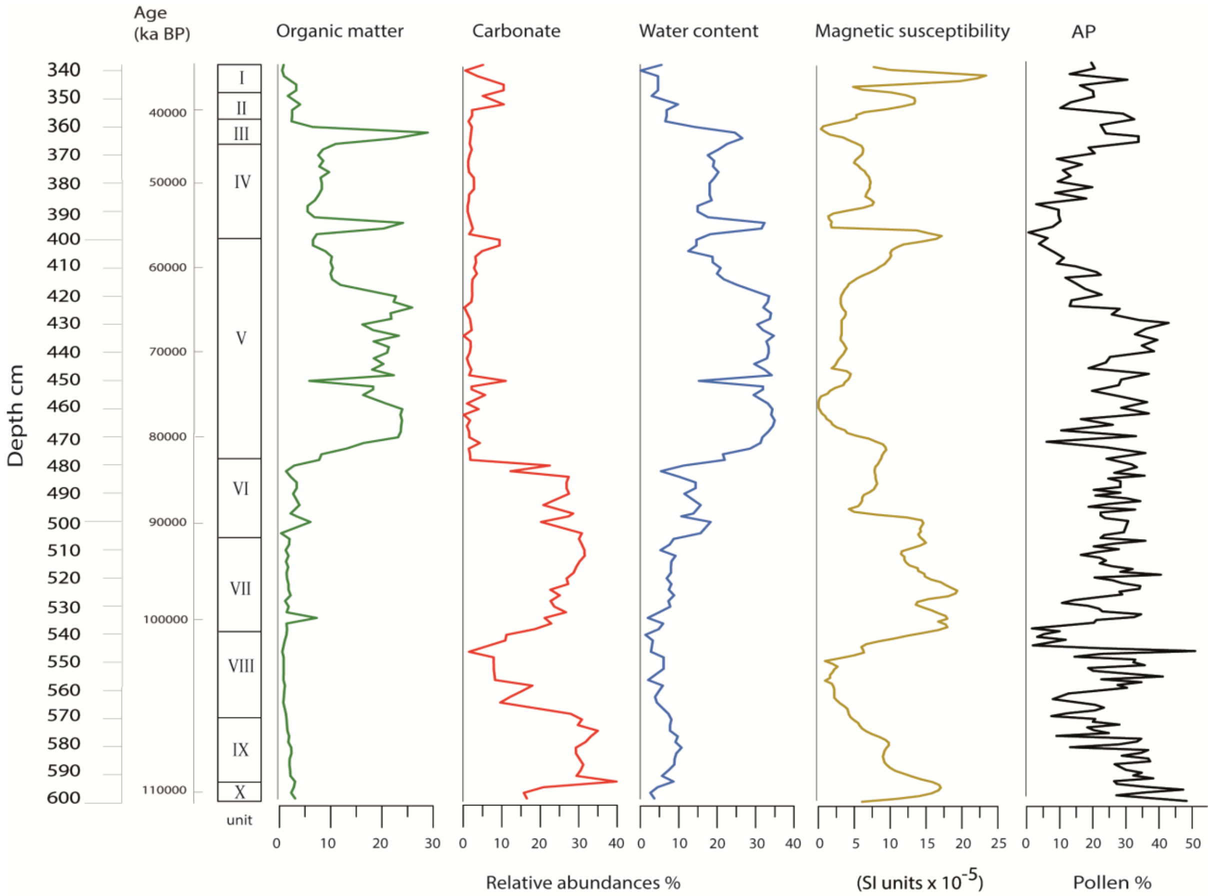

Figure 4.6 Percentage loss-on-ignition data showing organic matter, carbonate and water content of the Aammiq Wetland sediment core plotted against age and depth and differentiated into units to aid presentation of the results. Also displayed are variations in bulk magnetic susceptibility levels (κ_{LF}) and arboreal pollen (AP) percentages (Jeffers and Willis, in Press, Jeffers and Willis, submitted).

Units X, IX and VII of the LOI record shows high levels of carbonate material (≤ 40%). At the end of unit IX the abundance of carbonate decreases abruptly to less than 2% and remains low for the duration of unit VIII, before increasing rapidly to ≤ 35% at the start of unit VII. The abundance of carbonate declines during unit VI and then remains low (<

10%) for the duration of record. This decline is coincident with a rapid increase in organic matter content (> 40%). Organic matter content remains high until the later stages of unit V when it declines (11%). For the remainder of the sequence organic matter sedimentation remained below 20% with the exception of two spikes at the end of unit V (48%) and unit III (57%). The abundance of carbonate then increases slightly at the end of the sedimentary sequence during unit II and I.

4.6 Discussion

Over the sedimentary sequence a number of different processes can be inferred based on changes in geochemical composition, loss-on-ignition, magnetic susceptibility and sediment type. These processes include changes in landscape erosion, circulation of water through the karstic system, terrestrial and aquatic biological productivity and lake levels. The dynamic nature of this environmental system reflects, and was driven by, climatic variability.

The sedimentation of carbonate during a period of enhanced precipitation (Units X, IX and VII, ~ 112 to ~ 85 ka BP), corresponding to MIS 5e-MIS 5b.

The lower levels of the Aammiq sediment sequence (~ 112 to ~ 85 ka BP) had high concentrations of Ca (Fig. 4.5) and high levels of carbonate sedimentation (Fig. 4.4 and 4.6). The deposition of these lacustrine carbonates are typical of sediments in lake drainage basins where the surface sediments and soils are underlain by carbonate rocks (Kelts and Hsü, 1978, Alonso-Zarza, 2003), as are the Jebel Barouk-Jebel Niha Mountain adjacent to the wetland (Walley, 1996, Edgell, 1997). In these situations the lacustrine carbonates that develop primarily take the form of authigenic inorganic chemical precipitates, because of the high carbonate concentrations in the surface and groundwater (Gierlowski-Kordesch, 1998). We suggest the high levels of carbonate sedimentation at Aammiq reflects high rates of discharge from the karstified aquifer within the Jebel Barouk-Jebel Niha Massive, which is the main source of water input into the wetland (Walley, 1996), via the Nabba Aammiq contact spring

(Edgell, 1997). An influx of water with high carbonate content into the water body in the Bekaa Valley would have been derived from the dissolution of the limestone bedrock by the weak carbonic acid in rainwater. Palynological evidence (Jeffers and Willis, in press) showed high arboreal pollen abundance during this interval in time and thus the presence of a (mixed coniferous-deciduous) forest (Fig. 4.6) on the mountain slopes. The presence of forest would have enhanced the dissolution process as rainwater or snowmelt percolated through the slightly acidic soils and leaf litter layer that develops under a coniferous forest canopy (Hongve et al., 2000), thereby, slightly decreasing the pH of the water as it percolated through the bedrock (Develle et al., 2011).

The high level of carbonate production implies intense water circulation through the aquifer, and thus significant precipitation over the aquifer recharge zone on the mountain slopes. It is inferred that this level of water input and the need for open water to enable the precipitation of authigenic carbonates (Alonso-Zarza, 2003) indicates the persistence of a shallow lake in the Bekaa Valley. Similar high levels of lacustrine carbonate sedimentation during the last interglacial and early glacial period at Yammoûneh Basin, northern Lebanon (Develle et al., 2011) and Lake Orhid, Albania (Lezine et al., 2010) have also been shown to be associated with a period of enhanced precipitation, that led to an intensification of karstic activity and the development of an open water lake.

Low concentrations of Si, Al, Ti and K in the carbonate-rich sediment indicate that erosion from the mountain slopes during this period of high rainfall was low. The low levels of erosion supports the inference that the main source of carbonate sedimentation within the wetland was from authigenic precipitation within the water column rather than from the influx of detrital calcite. Low levels of fluvial erosion and transportation of siliclastic material into lakes where the catchment is primarily comprised of limestone is common because surface run-off is low due to the porosity of the bedrock (Platt and Wright, 1991). Erosion of the surface sediments would have been further reduced by the high levels of arboreal vegetation, that would have reduced erosion by raindrop impact and the propensity for surface run-off to occur (Castillo et al., 1997, Pimentel and Kounang, 1998). Low levels of organic matter sedimentation during this time of high terrestrial productivity, may reflect three possible processes: the stability of the catchment soils under a low erosional regime; organic matter entering the lake was rapidly decayed under high oxygen conditions (Meyers, 1994); or the incoming organic material was diluted by the high levels of carbonate sedimentation (Develle et al., 2011).

The sedimentary regime (low erosion-high carbonate) during the early glacial period also shows very high levels of MS (Fig. 4.6), which suggests that the magnetic grains were formed within the lake rather than transported to it. This supports the conclusion that catchment soils were stable during this time period. Calcium carbonate is diamagnetic with a negative susceptibility (Nolan et al., 1999) and therefore its presence in lake sediments significantly dilutes levels of MS (Dearing, 2001); therefore the magnetic minerals in the Aammiq lake sediments probably had a secondary biogenic origin (Lowenstam, 1981, Stolz et al., 1986) or were formed through authigenic/diagenetic processes (Hilton and Lishman, 1985, Anderson and Rippey, 1988), during the process of carbonate precipitation or post deposition.

Aeolian activity and the influx of sand, Unit VIII, (MIS 5d, ~ 108 to ~ 104 ka BP)

Carbonate sedimentation within the lake was interrupted by an influx of sand that started gradually at ~ 108 ka BP and ended abruptly at ~ 104 ka BP. It has previously been proposed by Develle et al. (2011) that the influx of quartz grains into the interior of Lebanon during glacial periods could be linked to either the global lowering of sea-levels that exposed more of the Mediterranean seabed, or derived from the Negev Desert, Israel approximately 300 km to the south of Lebanon. However, the presence of a Miocene alluvial fan on the lower slopes of the Jebel Barouk Massive (Walley, 1996), combined with the presence of coarse quartz grains (75-125 μm), and the significant quantities of limestone gravels (≤ 15 mm in diameter) in the sediment core imply a local rather than distal source for the quartz deposited into the lake. The increase in quartz sedimentation is associated in time with significantly higher concentrations of silica, of which quartz is one form (Trukhin et al., 2011). The corresponding decline in Ca concentration and low levels of carbonate production implies that the influx of quartz grains occurred during a decline in water circulation through the karstic system and thus precipitation over the mountain slopes. The influx of sand was initially in phase with a significant decline in AP abundance (~ 109 to ~ 106 ka BP), and thus forest cover (Jeffers and Willis, 2014), that would have resulted in a reduction in drag exerted by the forest canopy on airflow, increasing wind speeds and erosivity; thus the quartz grains for this period were likely transported by aeolian activity. Although sand continued to accumulate between ~ 106 to ~ 104 ka BP increased forest cover provides evidence that precipitation increased. Therefore it is likely that the influx of sand during this later period was derived from surges in surface run-off rather than aeolian activity.

The erosion of large quantities of siliclastic material from the watershed would generally be expected to result in high levels of MS (Thompson et al., 1975, Oldfield et al., 1978, Dearing et al., 1981, Oldfield et al., 1983); however, levels of MS are low consistently through the whole period that sand continued to accumulate. This episode of low MS may be due to the fact that quartz is diamagnetic (Hrouda and Kapička, 1986) and/or due to the effect of the sediment grain size. Grain size has been shown to significantly affect MS values of sediments (Thompson and Morton, 1979, Bradshaw and Thompson, 1985, Oldfield, 1991). The ferromagnetic component of the sediment, particularly if it is well sorted (as may be the case in an alluvial fan) would be concentrated in a particular range of grain sizes (Oldfield and Yu, 1994). As described above, the high levels of MS associated with the sedimentation of carbonates appears to show that at Aammiq the magnetic component of the sediment was primarily concentrated in small (silt/clay) sized particles.

Organic matter sedimentation and the development of a wetland (Unit V, MIS 5a-MIS4 (~ 84 to ~ 63 ka BP))

At ~ 84 ka BP there was an abrupt decline in carbonate production and a simultaneous increase in sedimentation of organic matter. The decline in carbonate is indicative of a reduction in karstic spring discharge, and thus lower precipitation over the recharge zone. The timing of this decrease in precipitation corresponds to a period of precession minima (84-78 ka BP), that results in maximum insolation in the northern hemisphere during the summer months (Lowe and Walker, 1997, Bradley, 1999). Whilst this may not have directly affected the annual level of precipitation, in the Aammiq region which primarily occurs during the winter (Storey, 2007), it would have increased summer temperatures and thus evaporation of water in the lake. As organic matter accumulated in the shallower waters, the inundated or saturated soils would have quickly become anaerobic as the respiring organisms consumed available dissolved oxygen in the water and trapped in the soils (Craft, 2002). Decomposition of the organic matter under these anaerobic conditions would have been reduced (Ponnamperuma, 1972) thereby slowing the degradation process and increasing organic matter accumulation rates. As levels of organic matter increased in the shallow waters, the lake transitioned into a wetland/swamp. A major source of the organic matter deposited in the wetland sediments may have been detritus from semi-aquatic grasses (e.g *Phragmites australis*, common reed), which the pollen spectra indicates had increased in abundance, at this time (Jeffers and Willis, submitted). The corresponding increase in concentrations of the element Fe was probably due to the development of redox conditions in the wetland, which is known to increase the mobility of this element (Ng and King, 2004, Willis et al., 1997). The increase in P may be linked to its absorption by Fe and Al or the removal of dissolved organic phosphorous from the water through plant uptake, absorption by soils and its subsequent incorporation into the sediments within the wetland (Curtis, 1985).

At ~ 74 ka BP, increased concentrations of Si, Al, Ti, K, Mg and Na in the lake sediments principally indicate the influx of minerals that had been eroded and weathered from the lake's catchment. This event was concurrent with a period of increasing abundance of arboreal vegetation, which suggests that rainfall levels had increased, and surface run-off was the primary factor causing weathering and erosion of surface sediments and exposed bedrock. This was another period of low carbonate production. If precipitation had increased sufficiently to facilitate spring water discharge into the wetland the low levels of carbonate production may have been due to carbonate suppression by respiration in living (and decay of dead) plants in the wetland. As increased respiration during the decay process increases the partial pressure of carbon dioxide (Pco_2), it lowers the pH of the water and thus inhibits the precipitation of calcite (Deocampo and Ashley, 1999).

Catchment instability and slope erosion (Units IV-I, MIS 3 (~ 60 to ~ 35 ka BP))

At ~ 60 ka BP, during the full glacial period an abrupt decline in organic matter sedimentation occurred that was coupled with an increase in the concentrations of siliclastic elements (Si, Ti and Al) in the sediments. There was also increase in clay particles and MS, which taken together are interpreted as indicating a significant reduction in organic productivity and an increase in weathering and erosion in the watershed. In the context of a landscape that was becoming increasingly treeless, as shown by the decline in AP, the decrease in organic matter probably represents a reduction in terrestrial and aquatic organic productivity, whereas the increase in erosion probably indicates an intensification of aeolian activity in an increasingly arid and open landscape. Further, declining Fe concentrations in the sediment suggest an increase in aerobic conditions in the wetland as water levels dropped and the sediments and soils became unsaturated and therefore aerated.

At the end of this arid period, there was a short-term humid event recorded in palaeoclimatic studies in Yammoûneh Basin, northern Lebanon (Develle et al., 2011, Gasse et al., 2011). At Aammiq wetland to the south, this event was indicated by the development of a peat layer that represented about 2000 years (~ 44 to ~ 42 ka BP) of higher water availability in the southern Lebanon Mountain range, as in the north. Wetter conditions favoured the growth of an oak and coniferous forest (*Cedrus libani*, *Pinus* and *Abies*) on the Lebanon Mountains (Jeffers and Willis submitted), and the sediment record suggests that terrestrial plant detritus contributed significantly to the accumulation of organic matter within the wetland at this time.

A further humid event from ~ 39 to ~ 35 ka BP is inferred from the increase in concentrations of Ca and carbonate and higher values of MS. This was concurrent with increased sedimentation of sand and limestone gravels into the wetland, which are believed to be derived from the alluvial fan that drains into the wetland. Taken together these results provide evidence for the recommencement of some water circulation through the aquifer and an increase in precipitation over the southern Lebanon Mountains during the later stages of MIS 3 prior to the onset of the last glacial maximum (LGM). This conclusion is supported by previous studies from the southern Levant (Israel). Lake levels of Lake Lisan (100 km south of Aammiq) the precursor to the Dead Sea, increased from ~ 40 ka BP (Lisker et al., 2010) attaining their highest level -164 m a.s.l just before the LGM at ~ 27 ka BP (Bartov et al., 2002, Frumkin et al., 2011). However, our interpretation is in contrast with reconstructions at Yammoûneh Basin that indicate an increasing local water deficit (Develle et al., 2011, Gasse et al., 2011). What could account for this north-south Lebanon difference in reconstructed water availability? Develle et al. (2011) suggest this may reflect water storage in glaciers on the higher slopes of Mount Lebanon, in the form of snow and in permafrost soils rather than a reduction in precipitation per se. The inferred higher levels of effective precipitation over the southern Lebanon Mountains may be due to their lower elevation relative to the northern section, and thus slightly higher temperatures. These would enable snow to melt and rainfall to infiltrate through un-frozen surface soils and sediments, and thus circulate through the karstic aquifer into the wetland.

4.7 Conclusion

At Aammiq Wetland, changes in sediment magnetic properties, geochemical composition, loss-on-ignition and sediment type record the effects of climatic and environmental events on wetland hydrology and structure at the end of the last interglacial and during the last glacial period.

At the end of the last interglacial and the start of the last glacial period (~ 112 to ~ 85 ka BP), there was a high level of water moving through the karstic system and into the wetland. This was generally a time of high forest cover and stable soils with associated low levels of fluvial erosion and transportation of siliclastic material into the lake. This period of long-term stability of the catchment soils was interrupted by periods of high erosion. The first of these (~ 108 to ~104 ka BP) was associated with high quantities of wind-blown clastic material into the lake. The second from ~ 84 ka BP led to the development of an anaerobic wetland/swamp that received greater input of fluvial derived sediments. There was then a major shift (from ~ 70 ka BP) within the catchment from high soil stability and vegetation cover to marked increases in erosion, decline in vegetation cover, and multiple changes in sediment type. This occurred against a backdrop of increasing aridity during the full glacial period. While palaeoclimatic reconstructions from northern Lebanon show increasing water deficit during the later periods of the glacial period, our data indicate the presence of increasing local water availability. That is assumed to derive from an increase in precipitation and thus water circulation through the karstic system into the wetland. The higher levels of water availability in southern Lebanon might well explain how this wetland and associated montane vegetation has managed to remain resilient to the climatic perturbations of the past 110,000 years.

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Chapter 5: Vegetation response to climate change during the last interglacial-last glacial transition in the southern Bekaa Valley, Lebanon.

5.1 Abstract

During the late-Quaternary extensive forests covered the slopes of the southern section of the Lebanon Mountains. Investigating the response of these forests to the climate change at the last interglacial-last glacial transition provides insight into their response to an abrupt and extensive deterioration in climate. Here we present the results of a new sedimentary sequence from the southern Bekaa Valley, Lebanon which spans the late last interglacial (MIS 5e) to early MIS 5c of the last glacial. High resolution analyses for pollen, isotopes of carbon, total organic carbon has been undertaken and reveals dramatic shifts in the vegetation composition during this transition. At the last interglacial (MIS 5e) to glacial (MIS 5d) transition an environmental threshold was crossed in response to a significant reduction in precipitation with the development of grassland/steppe dominated by Chenopodiaceae and *Artemisia*. The arid conditions lead to the loss of the temperate deciduous arboreal species (*Betula*, *Alnus* and *Corylus*) and significant reduction in the populations of coniferous trees *Cedrus libani* and *Pinus* that populated the slopes of the Lebanon Mountains during the last interglacial. At the onset of the MIS 5c interstadial an increase in arboreal pollen is observed in particular *Cedrus libani* as the grassland/steppe transitioned into an open forest-steppe. The response of the vegetation during the last interglacial to temperatures that were at least as high or higher than present, and the onset of the last glacial, when precipitation declined significantly has the potential to indicate how Mediterranean biotic communities will respond to changes in these climatic parameters that are predicted to occur in the near future.

Keywords: climate change, last interglacial-glacial transition, Cedrus libani, Lebanon

5.2 Introduction

In the future, Mediterranean mountainous ecosystems are predicted to be significantly affected by higher temperatures and reduced rainfall (IPCC, 2007). Vegetation reconstructions that span the last interglacial, when temperatures were at least as high or higher than at anytime during the Holocene (Kukla et al., 2002, Andersen et al., 2004, Felis et al., 2004), and the onset of the last glacial, when moisture availability declined significantly (Tzedakis, 2003, Muller and Kukla, 2004) can therefore be extremely useful for determining the response of Mediterranean ecosystems to both high temperatures and low rainfall.

During the last interglacial, as insolation levels started to decline at 122 ka (Berger et al., 2007), ice-sheets re-advanced across Fennoscandia and northern Europe. The impact of this ice-sheet growth on the climate of Mediterranean Europe was extensive; it altered oceanic (Rohling et al., 1998, Cacho et al., 2000) and atmospheric circulations (Kaspar et al., 2007, Enzel et al., 2008), sea (Sánchez Goñi et al., 1999, Bartov et al., 2003) and air (Allen et al., 1999, McGarry et al., 2004) temperatures and rainfall (Kallel et al., 2004, Almogi-Labin et al., 2009). Of all these climatic parameters variation in moisture availability between interglacial and glacial periods was the primary driving mechanism of Mediterranean forest dynamics (Magri et al., 2004).

Previous palynological studies conducted in the central Mediterranean region at Ioannina (Tzedakis, 1994) and Tenaghi Philippon (Wijmstra, 1969) Greece, Valle di Castiglione (Follieri et al., 1988) and Lago Grande di Monticchio (Brauer et al., 2007) Italy and Lake Orhid, Albania (Lezine et al., 2010) that span the last interglacial-last glacial boundary demonstrate the contraction of closed canopy forests that had persisted during the last interglacial and the expansion of steppic herbs in response to a steep reduction in moisture availability. However, the impact of this reduction in rainfall on Mediterranean vegetation communities in Bekaa Valley, Lebanon remains unknown.

This study presents the findings from a new high resolution terrestrial record spanning the last interglacial-glacial boundary from a sediment core obtained from Aammiq wetland in the Southern Bekaa Valley, Lebanon (Fig. 5.1). The key aims of this research are: i) to reconstruct the vegetation response to the climatic deterioration at the last interglacial-last glacial boundary, through analysis of pollen, stable isotopes of carbon and total organic carbon (TOC); and ii) to observe the forest dynamics during the late last interglacial warming (MIS 5e) in order to provide a potential scenario of biotic response to future climatic warming.

5.3 Study site

Aammiq Wetland is located in the southern Bekaa Valley, Lebanon (33°43'50''N, 35°47'10''E, 860 m a.s.l.) (Fig. 5.1). The wetland has a high international conservation status; it is designated as a RAMSAR site, UNESCO Man and Biosphere Reserve, and an Important Bird Area (IBA). To the east of the valley are the Lebanon Mountains and the lower slopes of the Jebel Barouk Mountain (1943 m) end at the wetland's edge. Parallel to this mountain range to the east of the valley are the Anti-Lebanon Mountains (highest peak Jabal el Shaykh 2814 m). The Bekaa Valley forms the central section of the Dead Sea Transform Fault (Butler et al., 1997). The valley sediments are comprised mainly of alluvial deposits consisting of silts, clays, sand and gravel that were laid down during the Quaternary Period (Walley, 1998).

The climate is semi-arid due to the rainshadow effect of the Lebanon Mountains, which reduces the moisture content of airflows from the Mediterranean Sea, thus creating a drier inland plateau (Sene et al., 1999). Average annual temperature in the southern Bekaa Valley is 16 °C (range 2-33 °C) and average precipitation is 600 mm yr^{-1}. On the higher mountain slopes average annual winter temperatures are below 0 °C and summer temperatures above 30 °C (Gasse et al., 2011). Precipitation on the mountain slopes that can

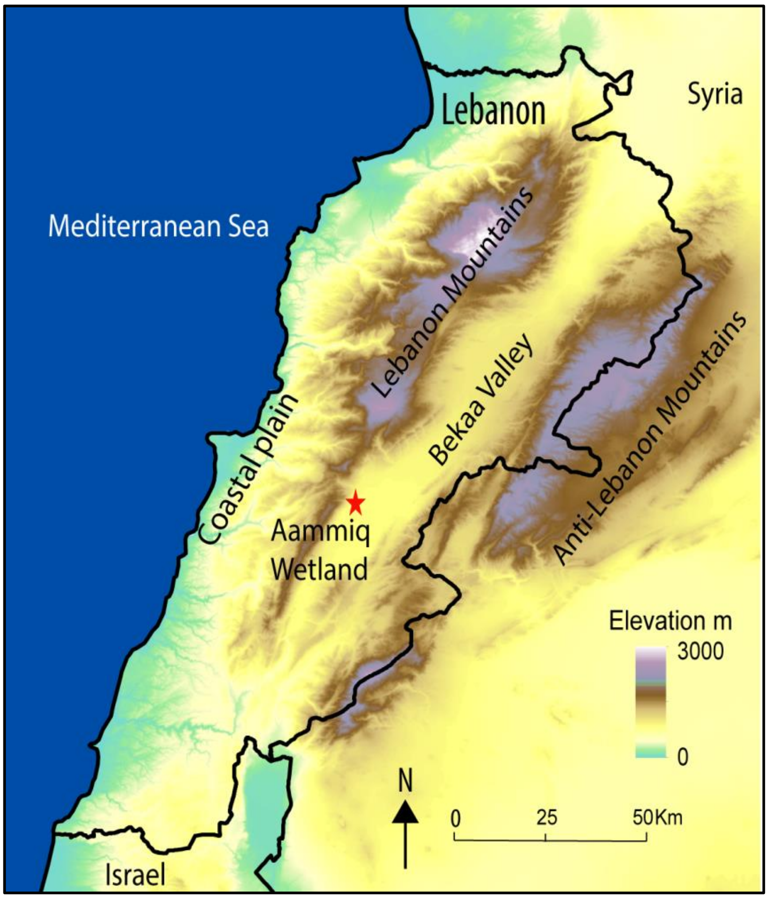

Figure 5.1 Topographic map of Lebanon showing Lebanon's four main topographical features (from west to east) coastal plain, Lebanon Mountains, Bekaa Valley and Anti-Lebanon Mountains. Also shown is the location of Aammiq Wetland.

reach 1600 mm yr^{-1} (Asmar, 2011) falls primarily as snow.

The wetland today encompasses > 70 ha during winter and spring (Storey, 2007), expanding up to 300 ha during high periods of rainfall or rapid snow melt (Storey, 2007). The wetland is the remnant of more extensive marshes and lakes that have been drained for agriculture (Walley, 1996; Storey, 2007) which is now the predominant land use across most of the Bekaa plain. The ephemeral wetland (due to over extraction of ground water) is a mosaic of different habitats including open water pools and reed beds comprised of *Phragmites australis* (common reed), *Typha spp.* (cattail/bulrush) and *Scirpus* (rushes). The eastern slopes and ridge of the Lebanon Mountains are covered by fragmented and largely open forests and scrub vegetation as the result of timber extraction. Remaining stands of trees include species of *Quercus calliprinos* (Palestine oak), *Q. infectoria* (gall oak), *Juniperus oxycedrus* (prickly juniper) and *Cedrus liban*i (cedar of Lebanon). These are predominantly located in the Al-Shouf Cedar Reserve situated within the Shouf UNESCO Biosphere reserve that covers the peak and upper slopes of the southern section of the Lebanon Mountains from Dahr Al-Baidar Mountain in the north to the Niha Mountain in the south.

5.4 Core extraction

A 7 m sediment core was obtained from Aammiq Wetland (33°43'51.47"N, 35°47'08.80"E) using a Russian corer in November 2008. The sediment core was wrapped in cellophane and aluminium foil to retain moisture during transportation and black plastic tubing to reduce exposure to sunlight so that Optically Stimulated Luminescence (OSL) dating could be conducted.

5.5 Dating

Chronology of the core was determined by a combination of radiocarbon dating (The ^{14}CHRONO Centre, Queen's University Belfast, UK) and OSL dating (Oxford Luminescence Laboratory, University of Oxford, UK). Radiocarbon dating was conducted on bulk sediments from two depths: 258 and 360 cm and sent to ^{14}CHRONO for sample preparation and Accelerator Mass Spectrometric analysis (AMS). In recognition of the potential age errors caused by the 'reservoir effect' of radiocarbon dates obtained from bulk sediments in a region where the bedrock is predominately comprised of limestone (Meadows, 2005) an attempt was also made to obtain radiocarbon dates from the pollen fraction of the sediment following the protocol by Brown et al. (1989). Unfortunately the pollen concentrations and hence quantity of carbon sent to^{14}CHRONO were too low to produce any dates.

Luminescence dating was performed at five depths: 219, 278, 486, 544 and 690 cm. In the laboratory, sediment (1 cm^3) was extracted from the centre of the core under subdued yellow laboratory lights. Quartz grains (75-210 µm) were extruded from the sediment (219, 278, 544 and 690 cm) using standard methods (Telfer and Thomas, 2007, Thrasher et al., 2009) which included 30% HCl and 30% H_2O_2 to remove carbonates and organics respectively. The remaining sediment was etched with 60% HF to remove feldspars and the outer layer of the quartz grain affected by alpha radiation. The quartz was then wet sieved to isolate grains 75-210 µm in size. Only fine quartz grains (5-15 µm) were retrievable from the sediment at 486 cm, and this was achieved by applying additional wet sieving and density separation through flotation in H_2O. Grains were analysed using the Single Aliquot Regenerative (SAR) protocol of Murray and Wintle (2000, 2003). Measurements were made on a Risø TL-DA-15 reader using blue diode illumination and U-340 filters.

Calibration of the radiocarbon dates (^{14}C ages) to years before present (cal. yr BP) was performed using CLAM's default northern hemisphere (atmospheric) calibration curve that uses IntCal09 (Reimer et al., 2009). The age-depth model combining the radiocarbon dates and luminescence dates was then constructed using linear interpolation in the software program CLAM version 2.1 (Blauuw, 2010) run in R programming language (R development core team, 2010).

5.6 Laboratory analyses

Pollen

Sub-sampling of the core was conducted at 1 cm resolution and 3 cm^3 of sediment from each level was removed for pollen analysis. Due to the problems associated with obtaining sufficient quantity and identifiable pollen grains from arid climatic regions (Horowitz, 1992), a pollen extraction method was devised that omitted both the application of alkali (e.g KOH, NaOH) which causes deterioration of pollen grains already exposed to oxidation (Woosley, 1978) and the acetylation (acetolysis) process that can degrade and destroy pollen (Hesse and Waha, 1989, Charman, 1992). Chemical treatment included 1) 7%

HCl at 90 °C for 60 minutes to remove carbonates 2) 60% HF at 90 °C for 90 minutes to remove silica 3) 7% HCl for 90 minutes to remove colloidal silica and silicofluorides created during the HF reaction.

To minimise loss of pollen from already low concentrations, decanting was reduced by 50% relative to a “standard preparation” pollen extraction method (Bennett and Willis, 2001) by using a fine sieving approach at (10 µm) over a vacuum flask. Using this method, humic acids and clays are removed and the sediment deflocculated without alkali or acetylation treatments and with minimal decanting. Extracted pollen was mounted in silicone oil. For levels 544-580 cm, > 300 terrestrial pollen grains were counted; for 580-600 cm, < 200 grains were recoverable due to extremely low pollen concentrations. Identification of the pollen was aided by the use of pollen identification keys (Faegri et al., 1989, Moore et al., 1991), pollen picture atlases (Reille, 1992, Reille, 1995, Reille, 1998) and referral to the pollen reference collection held within the Long-Term Ecology Lab.

TOC and $\delta^{13}C$

Sampling of the core for stable isotope analysis was conducted at 2 cm resolution and 1 cm^3 of sediment was extracted. Bulk sediments were decarbonized with HCl, homogenised and then dried. The samples were then analysed for total organic carbon (TOC) and isotopic composition of carbon $\delta^{13}C$ using a Costech Elemental Analyser attached to a Thermo MAT 253 mass spectrometer in continuous flow mode at the Godwin Laboratory for Palaeoclimate Research, Cambridge. Precision is < 0.1‰.

Data handling

The pollen was divided statistically into three pollen zones using optimal splitting by information content (Bennett, 1996) using the software PSIMPOLL version 4.27 (Bennet, 2009). The pollen percentages, pollen zones, TOC and $\delta^{13}C$ values were graphed using C2 software version 1.7.2 (Juggins, 2011).

5.7 Results

Dating

Results of the radiocarbon dates are presented in Table 5.1 and luminescence dates in Table 5.2. The results of the age-depth model are shown in Fig. 5.2.

Sample depth (cm)	Laboratory number	Sample Material	^{14}C age yr. BP	2σ Calibrated age (yr BP)
258	UBA-20316	Bulk sediment (Clay/organic rich)	12836 ± 69	14939-15882
360	UBA-20315	Bulk sediment (Clay/organic rich)	37205 ± 745	41058-43108

Table 5.1 Radiocarbon dating results from the Aammiq Wetland sedimentary sequence. ^{14}C ages were calibrated using IntCal09 (Reimer et al., 2009).

Sample depth (cm)	Lab code	Grains accepted to calculate D_e	Minimum Maximum grain size (μm)	K (%)	U (ppm)	Th (ppm)	Cosmic dose rate (Gy/ka)	Total dose Rate (Gy/ka)	D_e (Gy)	Age (ka)
690	OxL 2270	16	75-125	1.04 ± 0.05	4.06 ± 0.46	15.69 ± 1.56	0.10 ± 0.01	2.87 ± 0.18	355.56 ± 30.03	124 ± 13.2
544	OxL 2268	18	75-125	0.27 ± 0.007	1.91 ± 0.08	3.92 ± 0.25	0.12 ± 0.01	1.01 ± 0.05	105.6 ± 8.25	104.6 ± 9.5
486	OxL 2267	12 *	5-15	0.38 ± 0.02	3.4 ± 0.2	5.2 ± 0.26	0.10 ± 0.01	1.86 ± 0.17	163.0 ± 5.57	87.8 ± 8.6
278	OxL 2271	16	75-125	0.25 ± 0.12	3.4 ± 0.34	1.0 ± 0.1	0.16± 0.01	0.81 ± 0.05	16.28 ± 1.02	20.1 ± 1.7
219	OxL 2263	24	75-125	0.48 ± 0.02	4.52 ± 0.45	10.32 ± 1.0	0.18 ± 0.01	2.21 ± 0.15	11.38 ± 0.48	5.1± 0.4

Table 5.2 Results of optically stimulated luminescence dating. * D_e (equivalent dose) was calculated from 12 aliquots of fine grain quartz, assuming a moisture content of 15 ± 5%. Moisture content of all other samples was estimated to be 10 ± 3%.

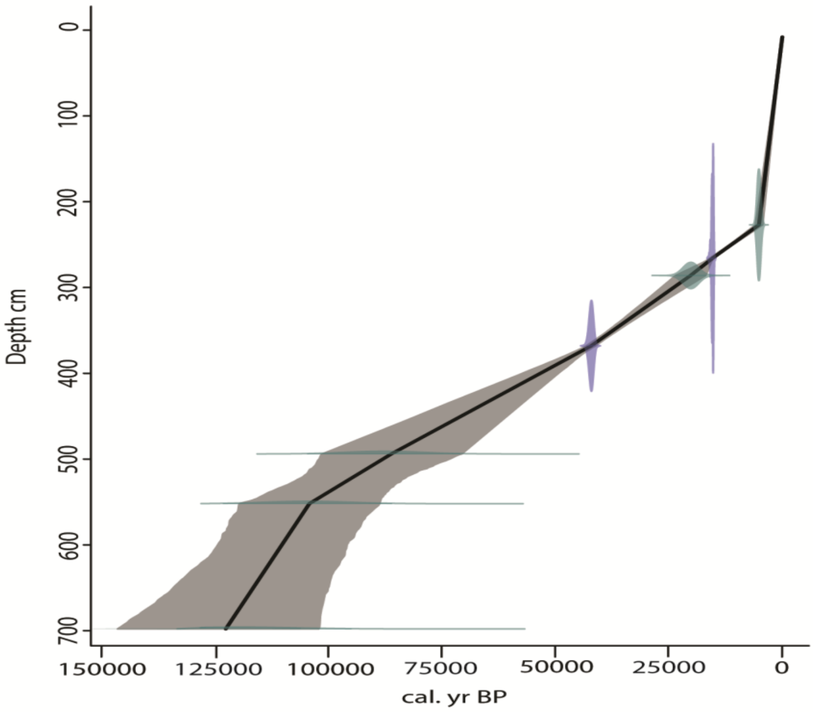

Figure 5.2 Age-depth graph undertaken in CLAM version 2.1 (Blaauw, 2010) using linear interpolation. Black line represents the 'Best fit' model and the grey shaded area 95% confidence limits. The probability distribution of the calibrated radiocarbon dates are shown as blue reduced plots, and the probability distribution of the OSL dates as green reduced plots.

Pollen

The results described here cover the section of core from 600 to 544 cm (Fig 5.3). The very low concentrations and degradation of the pollen grains from 600-700 cm prevented any reconstruction of the vegetation from this section of the sedimentary sequence. The results of the pollen analysis below 544 cm are currently unpublished data. In zone 1, which covers 600 cm to 585 cm (~112 to ~110 ka BP), the arboreal pollen (AP) ranged between 29 and 51% ($\bar{x}$ = 38%) of the total land pollen (TLP) (Fig. 5.3). The arboreal component included *Cedrus libani* (cedar of Lebanon), *Pinus* (pine), *Abies* (fir), *Quercus* (oak), *Alnus* (alder), *Betula* (birch) and *Corylus* (hazel). The non-arboreal pollen (NAP) assemblage was dominated by Poaceae (grass) and Compositae liguliflorae.

In zone 2, which covers 584 cm to 556 cm (~ 109 to ~ 106 ka BP), AP ranged between 10 to 40% ($\bar{x}$ = 27%) of TLP, which is lower than zone 1. There was total loss of *Alnus*, *Betula* and *Corylus* and a decline of *Cedrus libani* and *Pinus* to < 1% TLP and an increase in *Quercus*. There was a significant increase in NAP at this time and Chenopodiaceae (goosefoot) and *Artemisia* (daisy) dominated the landscape with ≤ 42 and 12% of TLP respectively.

At the onset of Zone 3 (~ 106 to ~ 104 ka BP), at 555 cm, AP was slightly higher and ranged between 25-43% ($\bar{x}$ = 30%) of TLP. There was a significant increase in *Cedrus libani* to ≤ 13% of TLP and an increase in *Pinus* to ≤ 6% of TLP. There was a compositional shift in the herbaceous pollen component with the decline in Chenopodiaceae to < 5% of TLP and an increase in Poaceae.

TOC and $\delta^{13}C$

Values of $\delta^{13}C$ were low (Fig. 5.3) and highly constrained (-26 to -28‰) from 600 to 544 cm (zone 1 through 3) with the exception of two spikes in values to -20‰ which occurred at 590 (zone 1) and 555 cm (between zones 2 and 3). Total organic carbon (TOC) values were low throughout the core (zone 1 through 3) ranging between < 0.1 to 0.7%. At 590 cm (zone 1) TOC increased to 0.5% where it remained relatively high (0.4 to 0.7%) until 557 cm when it started to decline to 0.1% at 544 cm (zone 1).

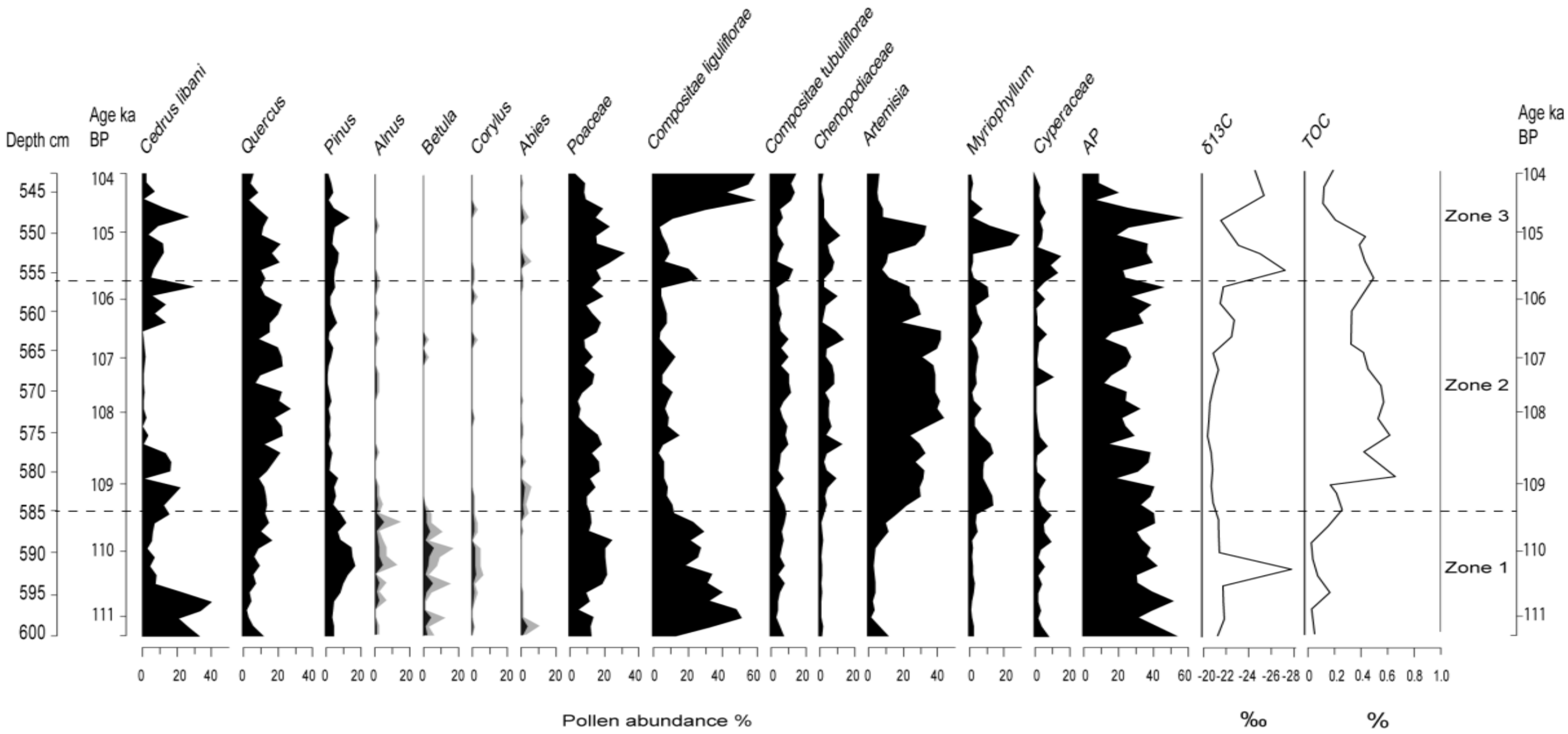

Figure 5.3 Pollen percentage diagram showing selected taxa and arboreal pollen (AP) for Aammiq Wetland constructed using C2 version 1.7.2 (Juggins, 2011) plotted against age and depth and divided into pollen zones. Exaggeration curves ×3 (grey shaded area) are displayed for *Alnus*, *Betula*, *Corylus* and *Abies*. Also displayed are stable isotopes of carbon ($\delta^{13}C$) and total organic carbon (TOC). The bottom of the pollen sequence corresponds to the later stage of marine isotope stage (MIS) 5e (~ 112 ka BP) and the top early MIS 5c (~ 104 ka BP).

5.8 Discussion

The termination of the last interglacial (MIS 5e, ~ 112 to ~ 110 ka BP)

The pollen assemblage from the lowermost section (600-585 cm) of this record corresponds to the very end of the last interglacial (~112 to ~ 110 ka BP) see (Fig. 5.3). During this time there was a significant presence of trees on the landscape as evidenced in the aboreal pollen (≤ 40 % of total land pollen). This included pollen from *Cedrus libani*, *Pinus*, *Abies*, *Quercus*, *Alnus*, *Betula* and *Corylus*. However, the pollen spectra also indicates that over 60% of the pollen rain was from non-arboreal types predominately Poaceae and Compositae liguliflorae. This pollen composition indicates that the landscape during this time was most probably mosaic in structure forming an open forest-steppe. The presence of temperate deciduous trees during the late last interglacial is coincident with the sedimentation of lacustrine carbonates at Aammiq that is inferred to be indicative of a period of high rainfall (Jeffers and Willis, submitted). It is likely that this enhanced rainfall would have facilitated the persistence of these deciduous species (e.g *Alnus*, *Betula* and *Corylus*) that are characteristic of temperate climates, are dependent on summer precipitation and cannot tolerate extended periods of drought (El-Moslimany, 1986, Cheddadi and Rossignol-Strick, 1995). Their presence during the last interglacial suggests that there was no marked seasonal moisture deficit, as exists at present in the eastern Mediterranean (EM) region (Kutiel et al., 1996, Ziv et al., 2006). A further consequence of the enhanced levels of rainfall may have been the down slope migration of the coniferous tree taxa, in particular *Cedrus libani* and *Abies* that today are limited to elevations above 1600 m a.s.l (Hajar, et al., 2008) due to their relatively high moisture requirements (Aussenac, 2002). The pollen percentages of these species may therefore not only reflect high levels of abundance on the upper mountain slopes but their presence in close proximity to the wetland thereby increasing the chance of their pollen grains becoming incorporated into the sediments.

Palaeoenvironmental observations made during past interglacials can provide a basis for comparison for the future evolution of the current interglacial period (Tzedakis et al., 2009) during which the Mediterranean (Gao and Giorgi, 2008, Giorgi and Lionello, 2008) and in particular the mountainous ecosystems are expected to be significantly affected by higher temperatures and reduced rainfall (IPCC, 2007). That temperatures were at least as high or higher than contemporary values (Kukla et al., 2002) during the last interglacial seems probable given that insolation levels at the onset of the last interglacial at 30˚N (latitude of southern Mediterranean) reached 535 Wm^{-2}, the highest value attained over the last 200,000 ka (Berger et al., 2007) and 11.58% higher than current insolation levels of 473 Wm^{-2}. The palaeoenvironmental evidence highlighted above from Aammiq suggests that these high temperatures were not coeval with a decline in rainfall (as future predictions for the current interglacial indicate) but were associated with an enhancement in rainfall that facilitated the observed expansion of forests across the Mediterranean basin (e.g Frogley et al. 1999, Tzedakis et al., 2002, Tzedakis et al., 2003, Brauer et al. 2007, Allen and Huntley, 2009, Gasse et al. 2011).

Accounting for such a disparity between the predicted future decline in rainfall with warming and the proxy records that indicate an increase in rainfall for the last interglacial is possible if the different primary mechanisms underlying the palaeorainfall record and future rainfall scenario are considered. The principal mechanism that facilitated high rainfall during the last interglacial in the EM was probably changes to synoptic scale weather systems that resulted in the increased frequency in the passage of summer depressions (Rohling and Hilgen, 1991). It is suggested by Rohling et al. (1998) that this increased frequency of the passage of rain bearing depressions mitigated the evaporative effect of higher temperatures during the last interglacial on soil water availability, in contrast with future scenarios that indicate higher temperatures are very likely to result in reduced summer soil moisture availability (Wang, 2005, Christensen et al., 2007), and a subsequent reduction in vegetation. This hypothesis is in general agreement with the conclusion reached by Arz et al. (2003) in a study of changes in Mediterranean moisture source during the Holocene humid period (9 to 7 ka BP), that regional circulation patterns controlling rainfall in the southeast Mediterranean and Near East may not be well simulated by regional climate models.

Last glacial inception (MIS 5d, ~ 110 to ~ 107 ka BP)

The transition from the last interglacial (MIS 5e) to the last glacial (MIS 5d) was initially marked by the loss of *Betula*, *Corylus* and *Alnus* and the increase of high altitude (1200-2000 m) coniferous species *Cedrus libani* and *Abies*. Despite the change in species composition, levels of AP remained relatively stable at this time. This change in the structure of the tree assemblage is concomitant with a compositional shift within the herbaceous community identified by the significant increase in Chenopodiaceae and *Artemisia*. The increase in these dry indicator species (Weinstein-Evron, 1983) with low precipitation requirements of 50-150 and 100-150 mm yr^{-1} respectively and the loss of the temperate deciduous trees (*Betula*, *Corylus* and *Alnus*) is indicative of a change in the seasonal distribution of rainfall. The influx of aeolian transported sand into the wetland from ~ 108 ka BP (Jeffers and Willis, submitted) is inferred to indicate that there was also a significant decline in the total annual rainfall volume.

The observed changes in moisture availability at Aammiq during the onset of MIS 5d were ultimately driven by changes in the Earth's orbit. The combination of the occurrence of the perihelion in the northern hemisphere winter combined with greater eccentricity and low obliquity resulted in the weakening of the northern hemisphere seasonal cycle of insolation. Consequently in the Mediterranean at the onset of MIS 5d at 109.50 ± 1.40 ka BP (Brauer et al., 2007) insolation levels at 30˚N during mid-June were 498 Wm^{-2} (a reduction of 6.9%) from the last interglacial maxima of 535 Wm^{-2}. In response to this global summer cooling there was a significant accumulation of perennial snow in the northern hemisphere (Vettoretti and Peltier, 2004) and consequently glacial advance at high latitudes. Initially the ice growth did not significantly affect oceanic heat transfers over the sub-polar north Atlantic (McManus et al., 2002), thereby enabling the continued flow of moisture into the Mediterranean and facilitating the persistence of tree populations (Tzedakis, 2005). At Aammiq this is observed in the significant continued presence of *Cedrus libani* and *Abies* as MIS 5d initially developed, although the seasonal distribution of rainfall in the EM must have been affected to account for the extirpation of the temperate deciduous tree species from the landscape. However, as the growth of ice sheets continued, it is probable that the oceanic heat transfer (thermohaline circulation) was disrupted resulting in the reduction of moisture availability for Mediterranean tree populations (Tzedakis, 2005). It was this reduction in moisture availability in the southern Bekaa Valley and the adjacent slopes of the Lebanon Mountains that probably facilitated the significant decline (< 1% of total land pollen) of the relatively moisture demanding populations of *Cedrus libani* during the later stages of MIS 5d.

The decline in *Cedrus libani* and *Pinus* as the stadial progressed was offset by an increase in *Quercus*. This is interpreted as indicating an increase in evergreen oak (*Quercus calliprinos*) that is native to the EM and effective at reducing water consumption when moisture availability is low (Salleo et al., 2002). The presence of evergreen oak species also suggests that average minimum winter temperatures on the lower slopes of the Lebanon Mountains and the Bekaa Valley during this stadial period remained above freezing, due to this type of oak trees intolerance of low temperatures for prolonged periods of time (Fady and Médail 2004). Unfortunately due to the degradation of the pollen grains in the sedimentary sequence no objective differentiation between evergreen and deciduous oak pollen types is possible. The concomitant increase in abundance of Chenopodiaceae (> 40% of total land pollen) and *Artemisia* (≤ 12%) indicates a climate with strong seasonality with cold dry winters and dry summers (El-Moslimany, 1987). The open steppe-forest that had persisted during the earlier stages of MIS 5d stadial was superseded by a dry herbaceous steppe.

The raised values of total organic carbon (TOC) (see Fig. 5.3) during this stadial is probably due to the increase in the growth of local communities of Chenopodiaceae. Given that a majority of species of Chenopodiaceae are halophytes (Davies and Fall, 2001) whose pollen is indistinguishable from the xerophytic form (Bottema, 1986) the increase in Chenopodiaceae may represent an increase in the halophytic type due to the expansion of suitable habitat as lake levels fell (Djamali et al., 2008) as well as its expansion across the wider landscape. The correspondence between Chenopodiaceae and TOC dynamics that was consistent for the entire stadial (MIS 5d) suggests that the primary source of TOC into the lake sediment was Chenopodiaceae detritus. This hypothesis is partially supported by the $\delta^{13}C$ record that displays values during MIS 5d of -25‰ to -27‰, indicating that the source of TOC was either C_3 terrestrial plant matter or lacustrine algae (Meyers, 1994). The presence of carbonate (alkali) rich sediment and shallow waters (Jeffers and Willis, submitted) would have provided ideal conditions for the observed expansion of aquatic *Myriophyllum* (watermilfoil) that thrives in inorganic sediment and relatively alkaline waters (Smith and Barko, 1990), usually in water depths of 1-4 m (Nichols and Shaw, 1986). That the increase in *Myriophyllum* was concurrent with the increase in Chenopodiaceae and the fact that the pollen curves of these two species followed similar trajectories throughout MIS 5d suggests that the palustrine system in the Bekaa Valley during this interstadial was comprised of a mosaic of open water pools and wetted soils.

Early MIS 5c (~ 106 to ~ 104 ka BP)

The onset of the interstadial (MIS 5c) in the southern Bekaa Valley is notable for the decline in Chenopodiaceae and the increase in AP. Within the forest mosaic there was a significant increase in coniferous species (*Cedrus libani* and to a lesser extent *Pinus* and

Abies). The re-expansion of the coniferous trees was probably facilitated by an abrupt and extensive increase in rainfall at the onset of the interstadial. A baseline for the minimum rainfall can be approximated at > 450 mm yr^{-1}, the minimum annual precipitation required for the persistence of *Cedrus libani* (Ducrey et al., 2008). The non-regeneration of temperate deciduous species (*Alnus*, *Betula* and *Corylus*) indicates the optimal climatic conditions for these species including a lack of summer seasonal moisture deficit as occurred during the late last interglacial was not reattained during this interstadial.

Comparisons with other regional palynological and palaeoclimate records

The vegetation record at Aammiq during the late last interglacial demonstrates some similarities but also differences with the only other pollen based vegetation reconstruction in the EM that covers this time period from the Yammoûneh basin in northern Lebanon (Gasse et al., 2011). At Yammoûneh significant levels of AP were also observed, however, the presence of significant levels of evergreen *Quercus* at Yammoûneh has been interpreted as being indicative of large seasonal contrasts in moisture availability (Gasse et al., 2011), for which there is no evidence at Aammiq. Although Yammoûneh (enclosed basin at 1360 m a.s.l) and Aammiq (open plain at 860 m a.s.l) are only separated by 50 km the different community structure is probably due to spatial heterogeneity in topography and hence difference in local climatic conditions. Vegetation reconstructions from the central Mediterranean at Ioannina in Greece (Tzedakis, 1994, Tzedakis, 2002, Frogley et al., 1999), Lago Grande di Monticchio in Italy (Brauer et al., 2007, Allen and Huntley, 2009) and Lake Orhid in Albania (Lezine et al., 2010) also display assemblages of AP but here the concentration is higher (> 70%) than Aammiq (≤ 40%) and taken to be indicative of the development of closed canopy forests around 127 ka BP that persisted (with a few brief periods of reduced forest cover) through the entire last interglacial (Tzedakis, 2003).

The increase in rainfall during the last interglacial documented at Aammiq is also recorded in the stable isotope spelothem record from Israel, where the depleted $\delta^{18}O$ values as interpreted by (Bar-Matthews et al., 2000, Bar-Matthews et al., 2003) indicate a period of high rainfall throughout the last interglacial. This enhancement of the rainfall regime is coeval with the deposition of sapropel S5 (Rossignol-Strick and Paterne, 1999) in the EM basin. The accumulation of these discrete organic rich layers in marine sediments (Cheddadi and Rossignol-Strick, 1995, Rossignol-Strick and Paterne, 1999) is thought to indicate that precipitation and associated run-off across the entire Mediterranean Basin was high and equal to or exceeded evaporation (Kallel et al., 2004).

At the onset of MIS 5d the significant expansion of steppic elements (Chenopodiaceae and *Artemisia*) observed at Aammiq is also recorded at Yammoûneh in northern Lebanon (Develle et al., 2011, Gasse et al., 2011). At Ioannina in Greece (Tzedakis et al., 2003) open vegetation expanded considerably and at Lake Ohrid in Albania (Lezine et al., 2010) and Lago Grande di Monticchio in Italy (Brauer et al., 2007, Allen and Huntley, 2009) there was a significant expansion of open vegetation and steppic herbs during the stadial (Melisey I). The record at Aammiq is also in general agreement with the record from Lake Urmia, NW Iran (Djamali et al., 2008) during the Espir stadial I (MIS 5d) that indicates the development of a grass steppe. The reduction in rainfall at the onset of MIS 5d that would have been required to facilitate the regional shift from forest/forest-steppe to steppe is observed as a rapid enrichment of $\delta^{18}O$ in the speleothems at Soreq (Bar-Matthews et al., 2000) and Peqiin caves, Israel, that are inferred to indicate a period of low rainfall (Bar-Matthews et al., 2003).

The expansion of forest as documented at Aammiq during the MIS 5c interstadial is also observed at Yammoûneh basin in northern Lebanon (Gasse et al., 2011) although the increase in AP is different in composition, at Aammiq being comprised predominately of coniferous tree taxa and Yammoûneh deciduous tree taxa. The record at Aammiq also demonstrates a similarity with the climatic and vegetation record at Lake Urmia, Iran during the Kaboudan interstadial I (MIS 5c) where there is a spike in AP that is similar in amplitude and magnitude to the AP peak of the last interglacial. However, the absence of *Zelkova carpinifolia* (a deciduous tree that requires spring and summer rainfall) suggests that although rainfall had increased significantly relative to MIS 5d, the climatic optimum (summer rainfall) of the last interglacial required for its expansion was not attained (Djamali et al., 2008). In the central Mediterranean at Ioannina (Tzedakis, 1994) and Tenaghi Philippon in Greece (Wijmstra, 1969) and Valle di Castiglione in Italy (Follieri et al., 1998) a significant increase in AP is also observed.

The observed increase in rainfall at Aammiq at this time is generally in agreement with climate reconstructions in Lebanon and Israel. At Yammoûneh, Lebanon the increase in $\delta^{18}O$ during MIS 5.3 (MIS 5c) is attributed to the development of wetter conditions that prevailed during MIS 5.4 (MIS 5d) (Gasse et al., 2011). In Israel the speleothem oxygen isotope records at Peqiin and Soreq caves documents low $\delta^{18}O$ events (-5‰ and -6 to -7‰ respectively) that are considered to be indicative of an increase in regional rainfall (Bar-Matthews et al., 2000, Bar-Matthews et al., 2003).

5.9 Conclusion

This palaeoenvironmental reconstruction provides the first independently dated record of terrestrial biotic responses to climatic change that occurred during the last interglacial-last glacial transition from the Bekaa Valley, Lebanon. The main findings of this study are:

1. At the end of the last interglacial (MIS 5e) the landscape mosaic was that of an open forest-steppe dominated by *Cedrus libani*. The presence of temperate deciduous species (*Betula*, *Alnus* and *Corylus*) in the forest matrix indicates that rainfall was likely higher than at present during the summer and prolonged intervals of aridity were absent. This suggests that the evaporative effect of higher temperatures during the last interglacial on summer soil

water availability for plant uptake was mitigated by an increase in rainfall volume. This observation is in contrast with future scenarios that indicate higher temperatures per se are very likely to result in reduced summer soil moisture availability.

2. The steep reduction in regional rainfall at the onset of MIS 5d facilitated the development of steppe dominated by low rainfall requirement/drought tolerant herbaceous species Chenopodiaceae and *Artemisia*. As climate within the stadial changed, the coniferous tree populations (*Cedrus libani* and *Pinus*) declined dramatically.

3. The transition to the interstadial MIS 5c is notable for the expansion of *Cedrus libani* and *Pinus* probably in response to an increase in rainfall. However, the non-rejuvenation of the temperate deciduous species that had persisted during the late last interglacial probably indicates that either the volume of rainfall or its seasonal distribution was not optimal for their re-colonisation of the Bekaa Valley.

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5.10 References

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Chapter 6: Eastern Mediterranean montane vegetation variability through the last glacial cycle-a record from the Bekaa Valley Lebanon.

6.1 Abstract

The climate of the last glacial period in the Mediterranean oscillated between dry and wet states within an overall trend of declining temperatures. As moisture availability is the primary control on plant species distributions in the Mediterranean, significant changes would be expected to be observed in plant community composition in response to this climatic variability. This paper investigates the response of the montane forests in southern Lebanon in the eastern Mediterranean to the climate variability of the last glacial period between ~ 104 to ~ 35 ka BP. Fossil pollen extracted from a new sedimentary sequence recovered from the southern Bekaa Valley, Lebanon shows that during marine isotope stage (MIS) 5c (~ 102 ka BP) an herbaceous steppe was replaced by coniferous-deciduous forest that was concurrent with an increase in regional precipitation. Twenty thousand years later (MIS 5a, ~ 82 to ~ 78 ka BP) precessional forcing of the climate system led to increased summer aridity, and a reduction in forest and expansion of grasses. Time series analysis of the pollen data showed that between MIS 5a to MIS 3 (~ 74 to ~ 40 ka BP) *Olea europapaea* (olive) and *Quercus spp.* (oak) expansion occurred at millennial intervals, in concurrence with Dansgaard-Oeschger (D-O) warming events. These cycles were overlaid upon a general declining trend in forest extent after ~ 63 ka BP, along with intense aridification, such that by ~ 54 ka BP, the landscape was almost totally devoid of trees. During the later stages of MIS 3 from ~ 45 ka BP as forest species recovered the steppic landscape was replaced by a forest-grassland.

Keywords: last glacial, montane forest, climate variability, southern Lebanon, Dansgaard-Oeschger events, precession minima.

6.2 Introduction

The climate regime of last glacial period is now known to be complex (Tzedakis et al., 2004). In the northern hemisphere the last glacial period was comprised of a succession of millennial-scale stadial (cold/dryer) and interstadial (warmer/wetter) periods (Rasmussen et al., 1997, Andersen et al., 2004). The shift between these two climatic states (stadial/interstadial) was the result of changes in orbital parameters (Grootes and Stuiver, 1997) internal climatic variability associated with Dansgaard-Oeschger (D-O) warming events (Dansgaard, 1993) and Heinrich cold events (Bond et al., 1992, Broecker et al., 1992). In response to this climatic variability there was a concurrent shift in the composition of vegetation communities.

During the late-Pleistocene glacial stadials the un-glaciated parts of the northern Europe the landscape was predominately covered by a treeless steppe-tundra (Van Andel and Tzedakis, 1996, West, 2000). During the interstadials, warmer and wetter conditions favoured the growth of open woodland (Woillard, 1978, Behre, 1989, Müller et al., 2003, Fletcher et al., 2010). In central and eastern Europe even during the coldest stages of the last glacial at least some of the landscape was populated by trees (Willis et al., 2000, Willis and Van Andel, 2004) that expanded during periods of climatic amelioration (Kuneš et al., 2008). In southern Europe, numerous palynological studies (Tzedakis, 1994, Allen et al., 1999, Sánchez Goñi et al., 2002, Tzedakis et al., 2004, Margari et al., 2009, Lezine et al., 2010) indicate that the vegetation mosaic shifted from being primarily comprised of steppic vegetation during cold/dry stadials to open forest/grassland during interstadials. In the eastern Mediterranean, terrestrial pollen records show a trend of forest decline as the last glacial period advanced, that was inter-dispersed by millennial-scale periods of forest expansion (Gasse et al., 2011).

There is now, therefore, a fairly good understand of vegetation dynamics in the western and central Mediterranean to climatic variability during the last full-glacial. What is still poorly known, however, due to the limited number of terrestrial vegetation records that cover all or part of the last glacial period, is responses in the eastern Mediterranean (Fletcher et al., 2010). Key questions that remain unanswered include did the vegetation communities in the eastern Mediterranean shift between the two alternate states of a forest/grassland and steppe in response to interstadial to stadial shifts in climatic conditions or did this region remain fully forested or steppic throughout? Was the eastern Mediterranean as has been inferred by Médail and Diadema (2009) an important area for refugial populations of tree taxa and if so, which trees? The aim of this study, therefore, was to address these knowledge gaps through analysis of fossil pollen from a new sedimentary sequence recovered from Aammiq Wetland, southern Bekaa Valley, Lebanon that covers the time period from the early last glacial period (~ 104 ka BP) to the late last glacial period (~ 35 ka BP) to examine the dynamics of the vegetation and the rates and nature of change over this important interval of climatic variability.

6.3 Study Area

Aammiq Wetland is located in the Southern Bekaa Valley, Lebanon (33°43'50"N, 35°47'10"E, 860 to 863 m a.s.l.) (Fig. 6.1). The wetland is designated as a RAMSAR site, UNESCO Man and Biosphere Reserve, and an Important Bird Area (IBA). With the exception of a few deep pools (Storey, 2007) the wetland today is ephemeral (Walley, 1996). During the late-winter spring the wetland typically encompasses ~ 70 ha that can temporarily expand to 300 ha during periods of heavy rainfall and snow melt (Storey, 2007). The modern wetland is a remnant of more extensive marshes and lakes that historical accounts suggest extended up to 30 km north of its current boundary (Storey, 2007). The extent of the wetland has contracted as land was drained and converted to agriculture and groundwater has been extracted for crop irrigation (Walley, 1996). The valley sediments are mainly comprised of alluvial deposits consisting of silts, clays, sand and gravel that were laid down during the

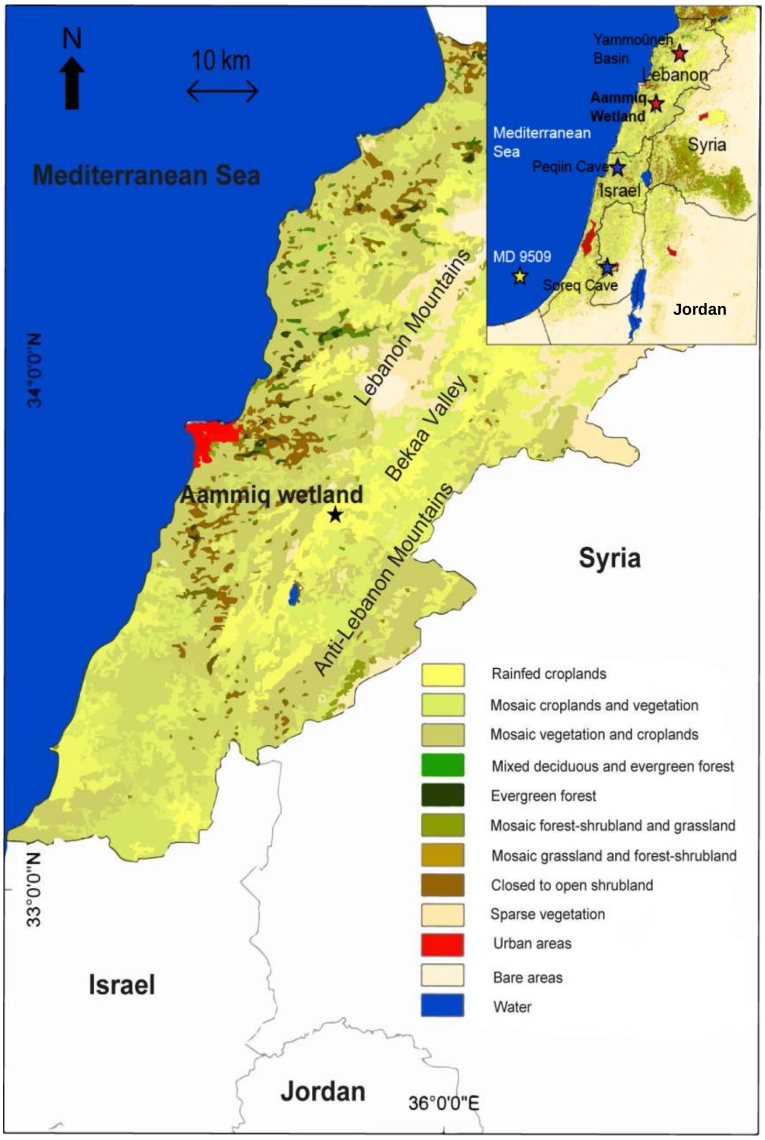

Figure 6.1 Land cover map of Lebanon constructed from GLOBCOVER information (Bicheron et al., 2009) showing location of Aammiq Wetland. Inset map shows location of other study sites mentioned in text.

Quaternary period (Walley, 1998). To the west of the valley lays the Jebel Barouk Mountain (1943 m) which forms the southern section of the Lebanon Mountain chain, the slopes of which end at the wetland's edge in the form of a Miocene alluvial fan (Walley, 1998) > 1.5 km in width. To the east of the valley lay the Anti-Lebanon Mountains (highest peak Jabal el Shaykh 2814 m).

The climate is Mediterranean with warm dry summers and cool wet winters. Temperatures in the Aammiq region of the Bekaa Valley range between 2 °C in winter to 34

°C in summer (Ministry of Environment/Med Wet Coast Project, 2004). On the higher mountain slopes average annual temperature of the coldest months (December, January and February) are below 0 °C and the warmest month (July) above 30 °C (Gasse et al., 2011). Due to the rainshadow effect of the Lebanon Mountains which reduces the moisture content of airflows that pass from the Mediterranean Sea over to the Bekaa Valley in the east (Sene et al., 1999), average precipitation is only 600 mm yr^{-1}. On the Lebanon Mountains, average precipitation on the upper slopes can reach 1600 mm year (Asmar, 2011). A majority of the precipitation falls from October to March in the Bekaa Valley as rainfall and on the mountains as snow.

The forests on the Lebanon Mountains are fragmented and largely open as a result of intensive timber extraction (Fady et al., 2008). The vegetation assemblages can be differentiated into three bioclimatic zones according to altitude. From the wetlands edge (860 m) to 1000 m a.s.l (Eu-Mediterranean zone) degraded *Quercus calliprinos* (Palestine oak) forest persists, between 1000 and 1600 m a.s.l (Supra-Mediterranean) the slopes are scattered with *Quercus infectoria* (gall oak) and *Juniperus oxycedrus* (prickly juniper) and above 1600 m a.s.l (Mountainous Mediterranean) stands of *Cedrus liban*i (cedar of Lebanon) occupy the mountain slopes (Hajar et al., 2008). The cedar forests are predominantly located in the Al-Shouf Cedar Reserve that covers the peak and upper slopes of the western flank of the southern section of the Lebanon Mountains. These forest remnants at all elevations are inter-dispersed by shrub vegetation with a predominance of the dwarf shrub *Sarcopoterium spinsum* (thorny burnet) that in arid and semi-arid regions of the Mediterranean is considered an indicator of land degradation (Ackermann et al., 2004).

Within the wetland, reed beds are the most abundant habitat (Conroy and Hepper, 2005) and are comprised of emergent plants (e.g *Phragmites australis*, common reed, *Typha spp*., reed-mace and *Scirpus*, rushes). The reed beds are inter-dispersed by open water pools up to depths of 2 to 3 m that contain over 200 floral species of submerged (e.g *Potamogeton spp.*, pondweed), and floating plants (e.g *Ranunculus spp.*, water crowfoot and duckweed, *Lemna.spp*) as well as algae *Chara spp.* (stonewort) (Conroy and Hepper, 2005).

6.4 Methods

Core extraction

A 7 m sediment core was extracted from within the wetland (33°43'51.47"N, 35°47'08.80"E) using a Russian corer in 2008. To reduce moisture loss prior to laboratory analysis the core was wrapped in cellophane and aluminium foil. To reduce the potential for exposure to sunlight so that optically stimulated luminescence (OSL) dating could be conducted the core was encased in black pvc tubing.

Dating

Chronology of the core was determined by a combination of radiocarbon dating conducted at ^{14}CHRONO Radiocarbon Laboratory, Queen's University Belfast, Belfast, UK and optically stimulated luminescence dating performed at the Oxford Luminescence Laboratory at the University of Oxford, Oxford, UK. Given that the greatest input of water into Aammiq wetland is from springs fed from kartisfied aquifers (Walley, 1996) it is highly likely that radiocarbon dates obtained either from bulk sediments or freshwater shells that obtain their carbonate entirely from dissolved inorganic carbon (DIC) are subject to the 'resevoir effect' of geologically old carbon entering the aquatic system, producing age estimates that are too old by several to many thousands of years (Meadows, 2005). An attempt to circumnavigate this effect was made by extracting the terrestrial fossil pollen fraction from the sediment for radiocarbon dating following the protocol of Brown et al. (1989). Unfortunately the concentrations of pollen sent to ^{14}CHRONO and hence carbon levels were too low to produce any signal (Hoper, pers comm., 2012). Bulk sediment samples for radiocarbon dating were therefore extracted from two depths: 258 and 360 cm and sent to ^{14}CHRONO for sample preparation and Accelerator Mass Spectrometric analysis (AMS).

Luminescence dating was performed at 5 depths: 219, 278, 486, 544, 690 cm. In the laboratory, 1 cm^3 of sediment from each depth was extracted from the core under subdued yellow laboratory lights. Quartz grains 75-210 µm in size were extracted from 219, 278, 544 and 690 cm following the standard method quartz grain extraction technique (Telfer and Thomas, 2007, Thrasher et al., 2009). The carbonate fraction of the sediment was removed using 30% HCl and the organic fraction using 30% H_2O_2. Feldspars and the outer layer of the quartz grain affected by alpha radiation were removed with 60% HF. The quartz was then wet sieved to isolate grains 75-210 µm in size. To isolate the fine grains (5-15 µm) from 486 cm the remaining quartz was wet sieved and density separated using flotation in H_2O. Grains were analysed using the Single Aliquot Regenerative (SAR) protocol of Murray and Wintle (2000, 2003). Measurements were made on a Risø TL-DA-15 reader using blue diode illumination and U-340 filters.

Calibration of the bulk sediment radiocarbon dates (^{14}C ages) to years before present (cal. yr BP) was performed using CLAM's default northern hemisphere (atmospheric) calibration curve that uses IntCal09 (Reimer et al., 2009). The Age-depth model for the sedimentary sequence combining the calibrated radiocarbon dates and the OSL dates was then constructed using linear interpolation in the software program CLAM version 2.1 (Blauuw, 2010) run in R programming language (2010).

Pollen

Sub-sampling of the core was conducted at 1 cm resolution from 544 to 486 cm and at 2 cm resolution from 486 to 338 cm. In the Laboratory, 3 cm^3 of sediment from each level was removed for pollen analysis. Due to the problems associated with obtaining a sufficient quantity of identifiable pollen grains from arid climatic regions (Horowitz, 1992), a pollen extraction method was devised that permitted the recovery of > 200 pollen grains from each level by minimising the loss of grains during chemical processing of the sediments. Relative to a "standard preparation" pollen extraction method (Bennett and Willis, 2001) this method omitted the application of alkali (used to deflocculate the sediment), acetylation (acetolysis, used to remove organics) and reduced decanting by > 50%. The method included 7% HCl to remove carbonates, 60% HF to remove silica and 7% HCl to remove colloidal silica and silicofluorides created by the HF reaction. The remaining sediment was fine sieved at (10 µm) over a vacuum flask to remove any remaining clays and humic acids and concentrate the organic fraction. This fraction was then dehydrated in Tert-butyl alcohol and mounted on slides in silicone oil. Pollen grains were identified at ×400 and ×1000 magnification using the pollen keys of (Reille, 1992) and (Moore et al., 1991) along with the reference collection in the Long-Term Ecology Laboratory, University of Oxford.

Numerical analysis

Zonation

The pollen data was divided statistically into six pollen zones using optimal splitting by information content (Bennett, 1996) using PSIMPOLL version 4.27 (Bennett, 2009). This was performed in order to identify sections (zones) of the fossil pollen assemblage in the sedimentary sequence that are different from adjacent sections (Birks, 1972) to enable easier description, comparison and correlation with other periods of time in the pollen sequence (Gordon and Birks, 1972).

Principle component analysis

Principle component analysis (PCA) was applied to the pollen data using C2 software version 1.7.2 (Juggins, 2011). When applied to large ecological data sets of interrelated variables, PCA can identify and display the primary directions of variance and patterns in the data (Davis, 1986). This is achieved by reducing the number of variables in a data set generally into two or three principle components (Webb et al., 1978), with the first axis indicating the direction of greatest variance and the second and subsequent axes capturing ever smaller amounts of variance (Bennett and Hicks, 2005).

Rarefaction analysis

The number of taxa in any pollen sample is dependent on the number of pollen grains counted and this varies between samples (Birks and Line, 1992). Rarefaction analysis compensates for differences in the number of pollen grains counted between samples by predicting what the number of pollen taxa would be if the pollen count for each sample was the same (Seppä and Bennett, 2003). Rarefaction analysis has the potential to provide insights into changes in taxonomic diversity and its potential causes (Seppä and Bennett, 2003). Rarefaction analysis was conducted using PSIMPOLL version 4.27 (Bennett, 2009). To standardise the pollen count of each sample the lowest pollen count of all terrestrial pollen (T_n=196) was used.

Rate-of-change analysis

Rate-of-change analysis (Grimm and Jacobson, 1992) was performed on the pollen data so that the rate of vegetation change over time could be estimated. The rate-of-change analysis that was applied using PSIMPOLL version 4.27 (Bennett, 2009), measures the difference between adjacent samples using chord distance and then divides the value obtained by the time lapse between the samples (Bennett and Humphry, 1995); the greater the value obtained the larger the rate-of change.

6.5 Results

Dating

Results of the radiocarbon and optically stimulated luminescence (OSL) dates are presented in Table 6.1 and 6.2 respectively and Figure 6.2. The time period covered by this paper is ~ 104 to ~ 35 ka BP, corresponding to 544-338 cm of the 7 m sediment core.

Sample depth (cm)	Laboratory number	Sample Material	^{14}C age yr. BP	2σ Calibrated age (yr BP)
258	UBA-20316	Bulk sediment (Clay/organic rich)	12836 ± 69	14939-15882
360	UBA-20315	Bulk sediment (Clay/organic rich)	37205 ± 745	41058-43108

Table 6.1 Radiocarbon dating results from the Aammiq Wetland sedimentary sequence. ^{14}C ages were calibrated using IntCal09 (Reimer et al., 2009).

Sample depth (cm)	Lab code	Grains accepted to calculate D_e	Minimum Maximum grain size (µm)	K (%)	U (ppm)	Th (ppm)	Cosmic dose rate (Gy/ka)	Total dose Rate (Gy/ka)	D_e (Gy)	Age (ka)
690	OxL 2270	16	75-125	1.04 ± 0.05	4.06 ± 0.46	15.69 ± 1.56	0.10 ± 0.01	2.87 ± 0.18	355.56 ± 30.03	124 ± 13.2
544	OxL 2268	18	75-125	0.27 ± 0.007	1.91 ± 0.08	3.92 ± 0.25	0.12 ± 0.01	1.01 ± 0.05	105.6 ± 8.25	104.6 ± 9.5
486	OxL2267	12 *	5-15	0.38 ± 0.02	3.4 ± 0.2	5.2 ± 0.26	0.10 ± 0.01	1.86 ± 0.17	163.0 ± 5.57	87.8 ± 8.6
278	OxL2271	16	75-125	0.25 ± 0.12	3.4 ± 0.34	1.0 ± 0.1	0.16± 0.01	0.81 ± 0.05	16.28 ± 1.02	20.1 ± 1.7
219	OxL 2263	24	75-125	0.48 ± 0.02	4.52 ± 0.45	10.32 ± 1.0	0.18 ± 0.01	2.21 ± 0.15	11.38 ± 0.48	5.1± 0.4

Table 6.2 Results of optically stimulated luminescence dating. * D_e (equivalent dose) was calculated from 12 aliquots of fine grain quartz, assuming a moisture content of 15 ± 5%. Moisture content of all other samples was estimated to be 10 ± 3%.

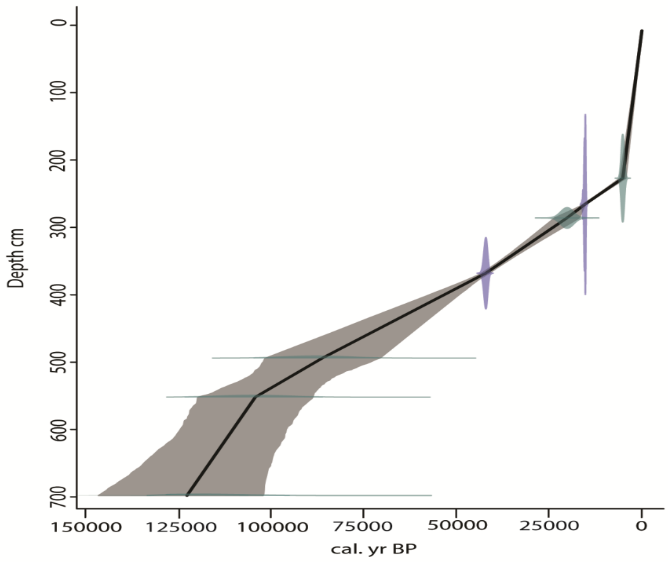

Figure 6.2 Age-depth graph constructed in CLAM version 2.1 (Blaauw, 2010) using linear interpolation. Black line represents the 'Best fit' model and the grey shaded area 95% confidence limits. The probability distribution of the calibrated radiocarbon dates are shown as blue reduced plots, and the probability distribution of the OSL dates as green reduced plots.

Pollen

Forty seven taxa of pollen were identified from the Aammiq Wetland sediment core. Taxa (15 in number) that are extremely rare have not been included in the pollen diagram (Fig. 6.3). The terrestrial taxa are expressed as percentages of total land pollen (TLP) and the aquatic plant pollen as percentages of the sum of the TLP and aquatic pollen.

In zone 6-1 (~ 104 to ~ 102 ka BP), which covers 544 to 534 cm the pollen sum is dominated by Compositae liguliflorae ≤ 60%. Other significant non-arboreal pollen (NAP) includes Compositae tubuliflorae and Dipsacaceae. The arboreal pollen (AP) < 25% ($\bar{x}$ = 11 %) is predominantly *Quercus*, with low percentages of *Cedrus libani*, *Abies* and *Olea europaea*.

In zone 6-2 (533 to 508 cm, ~ 101 to ~ 93 ka BP) AP increases but demonstrates a high degree of variability (< 13% and 43% total AP) ($\bar{x}$ = 27%). The main arboreal taxa are *Cedrus libani*, *Pinus*, *Abies* and *Quercus*. Between 530 and 520 cm *Alnus*, *Corylus*, *Tilia* and

Betula are recorded in the pollen record at very low values < 1%. There is also an increase in the herbs Umbelliferae and *Artemisia* at this time and a decrease in Compositae liguliflorae and Dipsacaceae.

In zone 6-3 (507 to 410 cm, ~ 92 to ~ 60 ka BP) AP increases very slightly ($\bar{x}$ = 29%). The main arboreal taxa are *Cedrus libani*, *Quercus* and *Pinus*. *Olea* appears in significant quantities (≤ 6%) from 480 cm. Poaceae values increase and fluctuate significantly (i.e between 6 and 66%). Other significant herbs include Chenopodiaceae, *Artemisia*, Umbelliferae, *Ephedra*, Compositae tubuliflorae, Caryophyllaceae, Rubiaceae, Leguminosae and Plantaginaceae. There is an increase in Cyperaceae and aquatic *Myriophyllum* and semi-aquatic Typhaceae that occurs infrequently and at very low levels throughout the zone.

In zone 6-4 (409 to 393 cm, ~ 59 to ~ 53 ka BP), the pollen assemblage is dominated by the herb Compositae tubuliflorae that increases from 8 to 64%. A significant decline in all other arboreal pollen and non-arboreal pollen with the exception of Dipsacaceae and Resedaceae is observed at this time.

Zone 6-5 (392 to 370 cm, ~ 53 to ~ 45 ka BP) is characterized by an increase in Poaceae to ≤ 30% and of Compositae tubuliflorae to > 25% and a decline of Compositae liguliflorae to 1%. There is also an increase in Chenopodiaceae, Caryophyllaceae, Ranunculaceae, *Centaurea* and Rubiaceae and the appearance of Lythraceae. Levels of AP remain below < 20 %.

In zone 6-6 (369 to 338 cm, ~ 45 to ~ 35 ka BP) there is an increase in all arboreal species (≤ 35%) with the exception of *Olea* that disappears from the pollen assemblage. There is a concurrent increase in the herbs *Artemisia* and Cyperaceae and aquatic *Myriophyllum* and Typhaceae as levels of Poaceae and Compositae tubuliflorae decline.

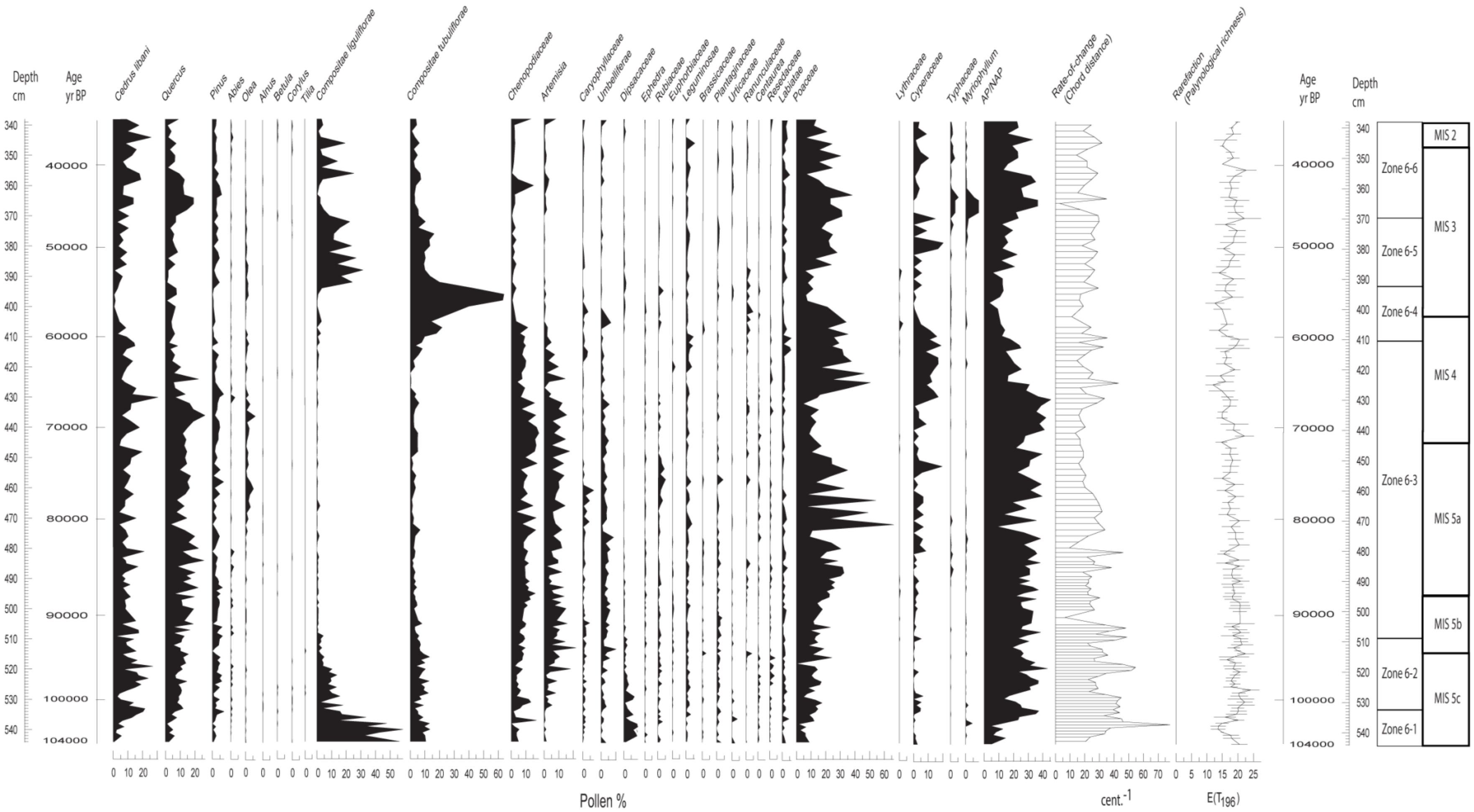

Figure 6.3 Pollen percentage diagram for Aammiq Wetland constructed using PSIMPOLL version 4.27 (Bennett, 2009) plotted against age and depth and divided into pollen zones. Rare pollen taxa are omitted from the diagram. The bottom of the pollen sequence dated at ~ 104 ka BP corresponds to marine isotope stage (MIS) 5 sub-stage c and the top the later stages of MIS 3 during the last glacial period (~ 35 ka BP). The diagram also shows the rate-of-change and palynological richness that takes all the observed pollen taxa (47) into consideration.

Rarefaction

The curve of the estimated palynological richness demonstrates a high degree of variability through time (12-27 taxa), Fig. 6.3. The low level of species richness that occurred during zone 6-1 (14 taxa) was concurrent with the decline in herbaceous taxa (Compositae liguliflorae, Cyperaceae, Rubiaceae, Cruciferae, Leguminosae and *Centaurea*). During zone 6-2 (≤ 27 taxa) and the start of 6-3 an increase in species richness was associated with the expansion of mixed coniferous (*Abies*) and deciduous (*Betula*, *Corylus*, *Alnus* and *Tilia*) tree taxa. Species richness in zone 6-3 declined (12 taxa) as *Abies*, Dipsacaceae, Ranunculacae, Euphorbiaceae and Brassicaceae disappeared from the pollen assemblage. At the zone 6-3 to 6-4 boundary palynological richness increased (≤ 22 taxa) due to greater diversity in the herbaceous community (e.g *Centaurea*, Ranunculaceae and Urticaceae). During zone 6-4 diversity declined (≤ 13 taxa) e.g Ranunculaceae and *Centaurea*. Zone 6-5 is characterized by an increase in species richness (≤ 20 taxa) due to the expansion of herbs (e.g Leguminosae, Labiatae and Dipsacaceae). Towards the end of the record (zone 6-6) species richness declined (15-20 taxa) as herb diversity was reduced (e.g *Ephedra*, Solanaceae and Euphorbiaceae).

Rate of change

The rate-of-change curve (Fig. 6.3) shows that the greatest rate of change occurred at the start of the record during zone 6-1 when the vegetation community transitioned from a herbaceous steppe with a predominance of Compositae liguliflorae to a forest/grassland. The rate-of-change during zone 6-2 is high, but displays significant variability between peaks and troughs, that are coincident with fluctuations in the abundance of forest taxa (i.e *Quercu*s, *Cedrus libani*, *Abies*). During zone 6-3 there was a decline in the rate of turnover in the plant community. However, this period was punctuated by an increase in the rate-of-change at ~ 63 ka BP as forest declined and grasses and herbs expanded (e.g Labiatae, Leguminosae and

Ranunculaceae). Through zones 6-4 to 6-6 the increase in turnover of species is concurrent with the expansion of tree taxa (*Cedrus libani* and *Quercus*).

PCA

Results of the PCA (Fig. 6.4) are displayed in the biplot of taxon vectors and sample scores. Axis 1 explains 39.56% of the total variance and primarily separates clusters of forest taxa (*Quercus* and *Cedrus libani*) and steppic herbs (Chenopodiaceae and *Artemisia*) from Compositae tubuliflorae and Compositae liguliflorae. Axis 2 explains 20.62% of the variance

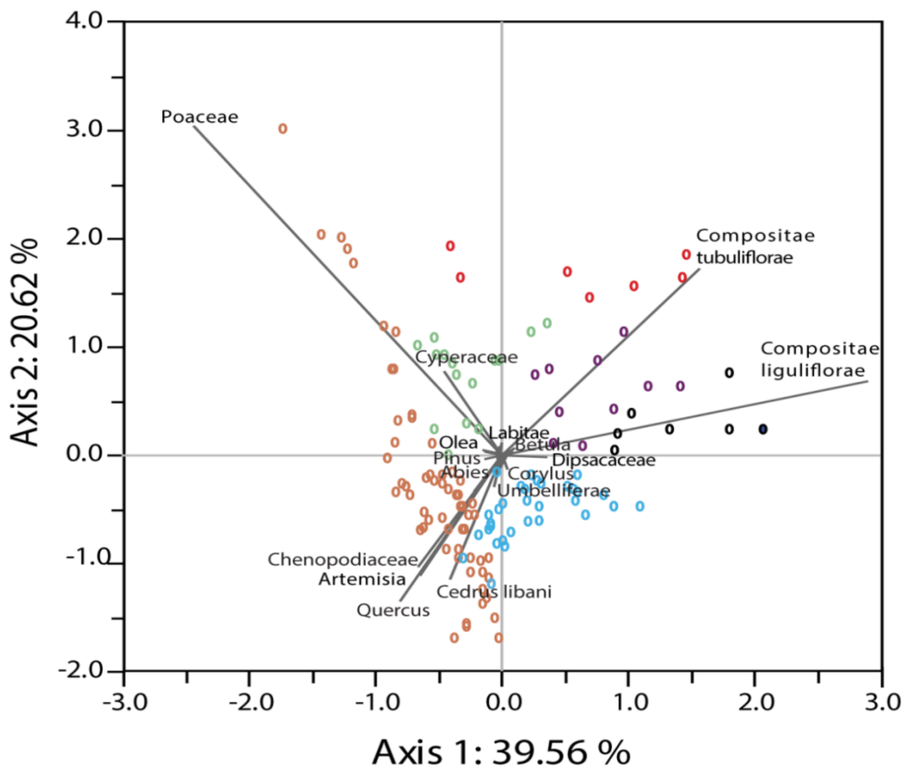

Figure 6.4 Biplot of PCA results for Aammiq Wetland pollen taxa showing the taxon vectors and sample scores. The pollen taxa (31) that account for very little of the variance are clustered at the intersection of the two axes and their labels have been omitted from the diagram. The sample score clusters (denoted by the coloured rings) correspond to the pollen zones 6-1 to 6-6 and thus different time periods: black 6-1 (~ 104 to ~ 102 ka BP), light blue 6-2 (~ 101 to ~ 93 ka BP), orange 6-3 (~ 92 to ~ 60 ka BP), red 6-4 (~ 59 to ~ 53 ka BP), purple 6-5 (~ 53 to ~ 45 ka BP) and light green 6-6 (~ 45 to ~ 35 ka BP).

and is dominated by the contrast between Poaceae and *Quercus*, *Cedrus libani*, Chenopodiaceae and *Artemisia*. Rare taxa are all clustered at the intersection of the two axes.

The sample scores show six clusters that correspond to the temporal differentiation between pollen zones (6-1 to 6-6).

6.6 Discussion

This study aimed to examine the relationship between climate change and vegetation variability in the southern Bekaa Valley and on the Lebanon Mountains during the last glacial period. Previous work has indicated that during the interval between 104-35 ka BP the climate in Lebanon Mountains alternated between intervals of low and high precipitation within a trend of declining temperatures. In the following section, the vegetation dynamics as reconstructed in this study are discussed within this climatic framework to examine the impact of these climatic events upon the vegetation of the southern Lebanon Mountains and the Bekaa Valley.

Forest expansion during the early glacial (MIS 5c to MIS 5b, ~ 104 to ~ 82 ka BP)

The main feature of the pollen record at the start of MIS 5c (from ~ 102 ka BP) is the decline in herbaceous pollen, in particular Compositae liguliflorae and the concurrent increase in *Quercus*, coniferous (*Cedrus libani*, *Pinus* and *Abies*) and with a temporal lag of approximately 1000 years, temperate (*Corylus*, *Alnus*, *Betula* and *Tilia*) and Mediterranean sclerophyllous *Olea europaea* L. var. *sylvestris* tree taxa. Palaeoclimatic records from Yammoûneh Basin, northern Lebanon (Develle et al., 2011, Gasse et al., 2011), Soreq and Peqiin Caves, Israel (Bar-Matthews et al., 2000, Bar-Matthews et al., 2003) show that there was a significant enhancement of rainfall at this time. We propose that rainfall levels were also high in the Aammiq region as indicated by the high levels of carbonate sedimentation into the lake in the Bekaa Valley, that reflects active circulation of water through the karstic system, and thus high levels of precipitation in the recharge zone over the slopes of the Lebanon Mountains (Jeffers and Willis, submitted a). This increase in water availability would appear to have favoured the expansion of tree taxa. The change in vegetation community composition is identified by the separation of the PCA sample score clusters 6-1 and 6-2 that distinguishes the steppe from forest/grassland that developed at this time (Fig. 6.4). The transition into forest also appears to have occurred at a relatively rapid rate (Fig. 6.3). Interestingly, this resulted in a lower level of taxonomic richness (Fig. 6.3) due to the temporary decline in the diversity of the herbaceous taxa (Cyperaceae, Rubiaceae, Cruciferae, Leguminosae and *Centaurea*) from the vegetation community and contrasts with other long-term ecological records from the Mediterranean (e.g Tzedakis et al., 2004, Margari et al., 2009) which indicate that at intervals of turnover resulting from climatic change, there is usually an increase in species richness due to combinations of old and new taxa co-occuring. The temporary extirpation of these herbaceous taxa in the Aammiq region that are generally light demanding (heliophilic), was likely due to the increase in canopy cover as the forest expanded, thereby reducing light penetration to the understorey vegetation.

At around ~ 103 ka BP there was deposition of sapropel S4 in the Mediterranean Basin (Martinson et al., 1987) that has been interpreted as indicating the persistence of pluvial conditions over the Mediterranean Sea (Kallel et al., 2000) and neighbouring continents (Rohling and Hilgen, 1991, Rohling and Thunell, 1999). As a consequence of the very high rainfall levels associated with this climatic event, the spatial distribution of trees within the forest/grassland mosaic in the Aammiq region would have differed significantly from the present day, where the extent of modern forests on the Lebanon Mountains increases along an elevational gradient as water availability increases (Hajar et al., 2008, Asmar, 2011). Results from this study indicate that precipitation during this interval in time appears to have been sufficient at lower elevations to have enabled *Abies* and *Cedrus libani* to grow around the lake basin at lower elevations; two species that today can only persist at elevations above 1600 m a.s.l (Hajar et al., 2008) due to their relatively high moisture requirements (Aussenac, 2002, Fontaine et al., 2007). It is probable that populations of *Cedrus libani* and *Abies* would have been interspersed with *Pinus* and *Quercus* throughout this period in time and with *Corylus*, *Betula*, *Tilia* and potentially even *Alnus* between ~ 100 and ~ 98 ka BP that would have been found in particularly moist areas like gullies or around the water bodies periphery (Margari et al., 2009).

The increased abundance of the herbs Chenopodiaceae and *Artemisia* from ~ 100 ka BP that are generally considered indicators of a dry climate (Weinstein-Evron, 1983, Weinstein-Evron, 1987, Rossignol-Strick, 1995) is concurrent with a reduction in rainfall in the eastern Mediterranean between ~ 100 to ~ 90 ka BP as observed by increasing $\delta^{18}O$ values in the Soreq Cave speleothems (Bar-Matthews et al., 2000). However, the continued persistence of forest in the Aammiq region, suggests that the increase in Chenopodiaceae and *Artemisia*, may more accurately indicate a highly seasonal climate with cold winters and dry summers rather than low rainfall per se (El-Moslimany, 1987). This would explain the presence of Chenopodiaceae and *Artemisia* in a landscape with significant tree cover and account for the similar trajectory of the PCA eigenvectors (Fig. 6.4) of these steppic herbs and cedar and oak. The presence of a strong seasonal climate may also account for the concomitant loss of temperate deciduous trees from the forests at ~ 98 ka BP. This is because temperate tree taxa (*Betula*, *Corylus*, *Tilia* and *Alnus*) cannot tolerate the prolonged periods of water limitation that are associated with a Mediterranean type summer climate (Cheddadi and Rossignol-Strick, 1995).

Development of grassland (MIS 5a, ~ 82 to ~ 78 ka BP)

During the period between ~ 82 to ~ 78 ka BP, *Quercus* declined and a grassland/steppe dominated by Poaceae became established on the landscape around the wetland. The timing of this change in state of the plant community (i.e ~ 82 to ~ 78 ka BP) corresponds to a period of precession minima, when maximum insolation in the northern hemisphere occurred during mid-June (Tuenter et al., 2005). Previous studies indicate that the first 2-2.5 ka of a period of precession minima is associated with an increase in aridity in the eastern Mediterranean (Tzedakis, 2007). As the climate became drier, reduced water availability appears to have favoured the expansion of grasses in southern Lebanon and also resulted in an increase in *Olea*, a tree species adapted to survive the aridity of the Mediterranean summer. However, the high abundance of *Cedrus libani*, which is more water-demanding than the Mediterranean oak taxa, suggests that the expansion of grasses and decline in oak trees may not have been driven by water availability. Instead, the decline in oak may have been a consequence of an increase in the frequency and intensity of fire activity (Jeffers and Willis, submitted b). The expansion of olive, which is highly light demanding (Rhizopoulou, 2007) may have been facilitated by high fire activity that created a more open landscape with fewer oak trees. During this arid period, the shallow and extensive root system of *Olea* (Spennemann, 1998) would have enabled it to extract moisture from the shallow soils on the lower mountain slopes or the Bekaa plain where it would have been restricted and probably provided a competitive advantage over other tree types.

The increase in Poaceae during this event is also concurrent with the change in state of the water body at Aammiq from a shallow open water lake to a wetland/swamp (Jeffers and Willis, submitted a). It is therefore probable that some of the increase in grass abundance was of semi-aquatic grasses (e.g. *Phragmites australis*, common reed), that is a primary species within the contemporary wetland (Conroy and Hepper, 2005). The increased abundances of Cyperaceae would have either colonised the wet soils in close proximity to the wetland/swamp (Genever et al., 2003, Caballero et al., 2005) or the open grassland (Duffin, 2008).

Millennial-scale forest expansion and contraction (MIS 5a-MIS 3, ~ 78 to ~ 40 ka BP)

The pollen data shows that between ~ 78 to ~ 64 ka BP the landscape on the slopes of the southern Lebanon Mountains reverted to a forest/grassland as *Cedrus libani* and *Quercus* (and to a lesser extent *Pinus* and *Olea*) increased in abundance. The observed increase in tree taxa in northern Lebanon (Gasse et al., 2011) at this time implies that the expansion of forest occurred across the breadth of the Lebanon Mountain range, in response to wetter climatic conditions (Gasse et al., 2011). It appears that during this time and continuing up to ~ 40 ka BP increases in the abundance of woody taxa was cyclical, occurring at the millennial-scale (Table. 6.3). The pacing of the expansion of tree taxa during the glacial period, in particular Mediterranean woody taxa (oak and olive) may be associated with Dansgarrd-Oeschger (D-O) climate warming events (in particular D-O events 20-14) recorded in the Greenland ice-cores (Dansgaard, 1993) and marine sediments as ice rafted debris events (Bond and Lotti, 1995).

Depth cm (Aammiq)	Age ka BP (Aammiq)	AP % directly preceding D-O event	AP % during D-O event	D-O event	Wolff et al., 2010 Age ka BP	Shackleton et al., 2004 Age ka BP	Svennson et al., 2008 Age ka BP
448	72.94	31	39	20	76.6	74	
440	70.13	28	41	19	72.28		
420	63.11	16	25	18	64.04		
413	60.66	14	25	17		60.14	59.44 ± 1.2
400	56.10	6	9	16		57.23	58.28 ± 1.2
394	53.99	10	13	15		53.52	55.8 ± 1.1
386	51.18	5	20	14		51.9	54.22 ± 1.1
374	46.97	13	19	13			49.28 ± 1.0
366	44.16	21	36	12		46.77	46.86 ± 0.9
358	41.52	25	35	11			43.34 ± 0.8

Table 6.3. Comparison of timing of D-O events 20-11 as determined by Wolff et al. (2010), Shackleton et al. (2004) and Svennson et al. (2008) and periods of arboreal pollen (AP) increase observed in the Aammiq Wetland pollen record.

Previous palaeoecological studies from the central Mediterranean in southern Italy (Allen et al., 1999), central Italy (Follieri et al., 1988, Follieri et al., 1998, Magri, 1999, Magri and Sadori, 1999) and Greece (Tzedakis, 1999, Tzedakis et al., 2002, Margari et al., 2007, Margari et al., 2009) all demonstrate millennial-scale pacing of forest expansion during the last glacial period that was coincident and strongest during D-O (interstadial) events 16-17 and 8 (Fletcher et al., 2010). In Lebanon, the greatest expansion of forest relative to the preceding time period occurred during D-O events 18-17 and 14 and also increased significantly during D-O events, 13-12. The intervals between D-O events were associated with the expansion of grasses, steppic (e.g Chenopodiaceae, Cruciferae and Caryophyllaceae) and non steppic herbs (e.g Resedaceae, Leguminosae and Dipsacaceae). This shift between forest and grassland/steppe vegetation communities was probably linked to a decline in water availability that was and continues to be the main factor driving forest expansion and contraction in the Mediterranean (Magri et al., 2004).

Within the context of D-O events, the observed expansion of forest during MIS 4 to its maximum extent during the glacial period at ~ 67 ka BP is significant because it was not coincident with a D-O interstadial event or with the expansion of *Quercus* or *Olea* but rather with an increase in coniferous trees (i.e *Cedrus libani*, *Pinus* and *Abies*). As detailed above, *Cedrus libani* and *Abies* have a high moisture requirement relatively to Mediterranean oak or olive, possibly suggesting that an increase in water availability was the driver of the expansion of coniferous tree populations.

A landscape devoid of trees (MIS 4 to MIS 3, ~ 64 to ~ 54 ka BP)

The rate-of-change within the vegetation community increased significantly at ~ *64* ka BP as the abundance of forest constituent taxa (*Quercus*, *Cedrus libani*, *Pinus* and *Olea*) declined and herbaceous taxa Poaceae, Cyperaceae, Rubiaceae and Leguminosae increased. The change in the composition of the vegetation community implies that the landscape changed from a forest/grassland to grassland/steppe. The initial contraction of forest was coincident in timing with Heinrich event 6 at ~ 65 ka BP that occurred during a cool period (67-59 ka BP) recorded in the GISP 2 ice core (Haase-Schramm et al., 2004). These melt-water forced Heinrich Events (Alley and Clark, 1999) reduced oceanic heat transport from lower to higher latitudes in the northern hemisphere (Chapman and Shackleton, 1998) that resulted in cooler and dryer conditions in the eastern Mediterranean (Langgut et al., 2011). The reduction in water availability and to a lesser extent the reduction in temperatures associated with H6 may have been the catalyst for the observed ecological change in the Aammiq region. However, given that the duration of H6 was < 1000 years (Langgut et al., 2011), the continued decline in forest was probably the result of a progressive decline in effective precipitation as the glacial period advanced (Gasse et al., 2011).

Whatever climatic factor initially drove the shift from a forest/grassland to grassland in southern Lebanon, the decline in forest continued into MIS 3, reaching its minimum extent at ~ 54 ka BP, when arboreal pollen accounted for less than 4% of the pollen sum. Thus, the full-glacial landscape during early MIS 3 appears to have been almost totally devoid of trees. The vegetation community on the open landscape was dominated by the light demanding (van Geel et al., 2011), xerophytic (Königsson and Poska, 1998) herbs of the group Compositae tubuliflorae, that accounts for low species richness at this time. The high abundance of this herb at this time could also be due to differential pollen preservation, as Compositae tubuliflorae is a pollen type that is resistant to oxidation (Cheddadi and Rossignol-Strick (1995). As shown in Figure 6.4, the sample cluster corresponding to pollen zone 6-4 distinguishes the Compositae dominated steppe that developed in place of the forest dominated by *Quercus* and *Cedrus libani* (pollen zone 6-3). The pollen record from marine core 9509 in the SE Mediterranean suggests that dry climatic conditions prevailed at this time, with aridification reaching its zenith at ~ 56 ka BP (Langgut et al., 2011).

Re-growth and persistence of forests (MIS 3, ~ 54 to ~ 35 ka BP)

The subsequent re-expansion of forest from ~ 54 ka BP was a gradual process, with arboreal pollen abundances remaining low (< 14 % total land pollen) until ~ 46 ka BP. Initially the slight increase in trees is coincident with an enhancement of rainfall (55-52 ka

BP) in the eastern Mediterranean (Bar-Matthews et al., 2000), that in the Aammiq region was sufficient to facilitate the erosion of silt and clay particles from the mountain slopes (Jeffers and Willis, submitted a). This increased rainfall was coincident with Dansgaard-Oeschger events 15 and 14 (~ 56 to ~ 51 ka BP). Higher summer temperatures in the Levant associated with this event (Ayalon et al., 2013) may explain why tree populations only expanded slightly; as precipitation levels may have been tempered by higher evaporation rates. Between ~ 45 to ~ 40 ka BP the forest once again expanded around Aammiq Wetland. Palaeoclimatic studies in Lebanon based on the sedimentation of peat (Jeffers and Willis, submitted a), $\delta^{18}O$ of ostracod valves (Gasse et al., 2011) and deposition of authigenic carbonates (Develle et al., 2011) indicate that this was a period of increased humidity. This climatic amelioration appears to have primarily favoured an increase in *Quercus* and to a lesser extent *Cedrus libani*, with no immediate response of *Pinus* to the climatic change (Fig. 6.3). In the wetland the increase in Typhaceae and *Myriophyllum* imply an increase in wetland habitat and the presence of some open water pools. The decline in Cyperaceae that is counterintuitive to the expansion of wetland habitat may be associated with the expansion of Typhaceae that is a dominant competitor in wetlands, excluding other plants with the growth of their dense canopy (Keddy, 2010).

During the later stages of MIS stage 3 (~ 40 to ~ 35 ka BP), the abundance of *Cedrus libani*, *Abies* and *Betula* expanded and *Quercus* and *Olea* declined. The change in forest community composition was probably due to the decline in temperatures in the Lebanon Mountains during the later stage of the last glacial period that were up to 10 °C lower than present day (Develle et al., 2011). Due to the fact that Olive (Moriondo et al., 2008) and Mediterranean oak are affected by low temperatures that persist for prolonged periods of time (Terradas, 1999), and *Cedrus libani* (Dirr et al., 1993), *Abies* (Aussenac, 2002) and *Betula* are extremely cold-tolerant (Aalto, 2006). The presence of significant abundances of coniferous and deciduous trees on the slopes of the southern Lebanon Mountains at this time is in contrast to the northern section of the Lebanon Mountains, where the landscape was almost devoid of vegetation (Develle et al., 2011). The dissimilarity between the pollen data recovered from Aammiq Wetland and Yammoûneh Basin separated by only 50 km, is possibly a consequence of the higher altitude of Yammoûneh Basin (1360 m a.s.l) relative to Aammiq Wetland (860-863 m a.s.l). Just taking the environmental lapse rate of 0.6 °C/100 m into consideration, temperatures in the Bekaa Valley were probably at least 3 °C higher than the Yammoûneh Basin, where the low temperatures (up to 10 °C lower than present) potentially facilitated the storage of water as ice in glaciers, as snow and in frozen soils (Develle et al., 2011). The landscape around the Yammoûneh Basin would probably have therefore resembled a periglacial landscape during this time. In contrast, the presence of forests on the lower slopes of the Lebanon Mountains in southern Lebanon, suggests that the soils would have needed to be unfrozen. The presence of *Cedrus libani* pollen (≤ 9%), a pollen type that is not transported long distances (Hajar et al., 2008), is for example, interpreted as indicating that populations of *Cedrus libani* were present on the southern slopes of the Lebanon Mountains and potentially in the southern Bekaa Valley at the end of MIS 3 prior to the onset of the last glacial maximum. It has previously been hypothesised by Fady et al. (2008) that Lebanon was a refugia for *Cedrus libani* during the last full glacial period; results from this study add further evidence to support this hypothesis.

6.7 Conclusion

The composition of the vegetation community in the southern Bekaa Valley and on the adjacent slopes of the Lebanon Mountains during the last glacial period varied significantly in response to climatic variability, primarily precipitation. At the start of MIS 5c (~ 102 ka BP) the herbaceous steppe dominated by Compositae liguliflorae was replaced by a mixed deciduous-coniferous forest/grassland that persisted through to MIS 5a (~ 82 ka BP) in response to a significant increase in precipitation.

During MIS 5a between ~ 82 to ~ 78 ka BP, the abundance of oak declined and a grassland developed. The expansion of grasses was coincident with an increase in summer aridity associated with a period of precession minima. However, the decline in oak may have been attributed to increased fire disturbance more than aridity. Whether the driver was aridity or fire (or a combination) the result was a more open landscape that facilitated an increase in the abundance of light demanding olive. The millennial-scale pacing of Mediterranean woody taxa expansion during MIS 5 through to MIS 3 (~ 40 ka BP) was coincident and probably associated with D-O climatic warming events. Between most of these D-O events forests contracted and grasses and herbs expanded.

Forest expansion reached its maximum extent during the glacial period at ~ 67 ka BP during a period of high rainfall. However, as the full glacial conditions began forest cover contracted until ~ 54 ka BP when the landscape was nearly devoid of trees. Tree taxa abundances remained low throughout the middle period of MIS 3 before re-expanding slightly during a period of increased humidity between ~ 45 to ~ 38 ka BP. As MIS 3 came to a close (~ 35 ka BP), effective precipitation was still sufficient to support a grassland/forest on the southern section of the Lebanon Mountains despite substantial declines in temperatures. The continued presence of *Cedrus libani* throughout this 70 ka vegetation record demonstrates the importance of southern Lebanon as a climatic refuge for this emblematic species through all of the last glacial period.

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6.8 References

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Chapter 7: Orbital and millennial-scale forcing of fire activity in Lebanon during the last glacial period

7.1 Abstract

This paper explores the relationship between local and regional fire activity, vegetation variability and climate change during the last glacial period ~ 104 to ~ 35 ka BP in the eastern Mediterranean through analysis of the charcoal and pollen records from a sedimentary sequence recovered from Aammiq Wetland, Bekaa Valley, Lebanon. Time series analysis of the macrofossil charcoal data identifies a periodicity of ~ 21 ka in peak local fire activity (at ~ 82, ~ 58 and ~ 39 ka BP) that is closely coupled to periods of precession minima. At ~ 82 to ~ 78 ka BP precessional forcing of the climate system resulted in a period of summer 'thermal maximum' that appears to correspond to, and is the probable driver of, the highest peaks in local and regional fire activity during the glacial period. The apparent impacts of these fire events on vegetation were reduced abundance of oak, and creation of a more open environment that was subsequently colonised by shade intolerant olive. Within these long term cycles, between MIS 5 and 3 regional fire activity varied over the millennial-scale with peak values of regional fire activity corresponding in time with Dansgaard-Oeschger (D-O) warming events, while low fire activity occurred when precipitation was high (~ 104 to ~ 82 ka BP) and temperatures were cool (~ 38 to ~ 35 ka BP).

Keywords: fire, macro and microcharcoal, Lebanon, last glacial period, precession minima, Dansgaard-Oeschger events.

7.2 Introduction

Wildfires are a naturally occurring phenomenon in the Mediterranean Basin where the incidence of fire is strongly linked to the prevailing climatic regime (de Luís et al., 2001). Due to the occurrence of high temperatures and low moisture that coincide during the summer months (di Castri et al., 1981) the flammability of biomass and propensity for fire to occur increases significantly (Gill et al., 1978). As a consequence fire is a significant natural driver of ecosystem variability in the Mediterranean (Arnan et al., 2006, Pausas, 2006), influencing vegetation dynamics over decadal to centennial (Colombaroli et al., 2007) and millennial (Vanniere et al., 2011) time-scales.

In response to fire, individual plants and populations (Pausas et al., 2004) employ one of two main strategies (Pausas and Vallejo, 1999, Pausas and Verdu, 2005). Some individual plants have the potential to persist due to a resprouting capacity (Bellingham and Sparrow, 2000, Bond and Midgley, 2001) a mechanism used by Mediterranean evergreen oaks (Pausas et al., 2008). In contrast some plant populations can persist by having a seed bank of hard coated fire resistant seeds (Naveh, 1975). The latter group may be stimulated to germinate by exposure to moderate heat or smoke (Staden et al., 2000), a common response in angiosperm seedlings (Pausas and Keeley, 2009) or survive by the long term storage of seeds within the canopy (serotiny), a strategy used by some Mediterranean pines (Lamont et al., 1991, Nathan et al., 1999) that are fire induced to disperse (Paula and Pausas, 2006).

The effect of changes to the fire regime on individual species populations and entire ecosystems in the Mediterranean during the Holocene has been well documented e.g (Carríon et al., 2003, Colombaroli et al., 2007, Colombaroli et al., 2008, Maja, 2007, Sadori and Giardini, 2007, Turner et al., 2008, Vanniere et al., 2008, Gil-Romera et al., 2010) and for an overview see Vanniere et al. (2011). Analysis of these circum-Mediterranean charcoal data indicates that fire activity and thus vegetation dynamics have been sensitive to changes in orbital parameters and millennial-scale forcing of the climate system (Vanniere et al., 2011) during the mid-Holocene (8500-2500 cal. BP), a phenomenon that has also been observed during the last glacial interglacial transition (Turner et al., 2008, Turner et al., 2010), and the full glacial period (Magri, 2008). Whilst these studies have demonstrated how fire activity varied in the Mediterranean Basin, due to the uneven temporal and spatial distribution of charcoal records there is still a poor understanding of the relationship between fire activity, vegetation and climatic variability in the eastern Mediterranean during the last glacial period. The aim of this study, therefore, was to address these knowledge gaps. We achieved this by analysing the temporal variations between orbital and millennial-scale climatic variability, fire frequency and intensity, and vegetation change during the last glacial period in the Aammiq wetland region of Lebanon.

This study focuses on analysing the macro and microfossil charcoal record from a high resolution sedimentary sequence covering the last glacial period from ~ 104 to ~ 35 ka BP from Aammiq Wetland in the southern Bekaa Valley, Lebanon. We assess the role of climate in altering fire regime variability and determine if any trends in fire activity exist and their potential relationship with orbital or millennial-scale climate forcing. Additionally, by comparing the charcoal data with the pollen from the same sedimentary sequence we examine the role of fire as an ecological disturbance mechanism and the potential role vegetation type and abundance has on fire frequency and intensity.

7.3 Study site

Aammiq Wetland is located in the southern Bekaa Valley, Lebanon (33°43'50"N, 35°47'10"E) at 860 meters above sea level (Fig. 7.1). It is the remnant of more extensive open water lakes and wetlands that once occupied a significant area of the Bekaa Valley (Walley, 1996, Storey, 2007). To the west of the valley is the Lebanon Mountain Range, with the lower slopes of the Jebel Barouk Ridge (1943 m) ending at the wetlands western periphery.

To the east of the valley are the Anti-Lebanon Mountains (highest peak Jabal el Shaykh 2814 m). The valley sediments are comprised mainly of alluvial deposits consisting of silts, clays, sand and gravel, that probably indicate multiple episodes of lacustrine and wetland sediment deposition during the Quaternary period (Walley, 1996).

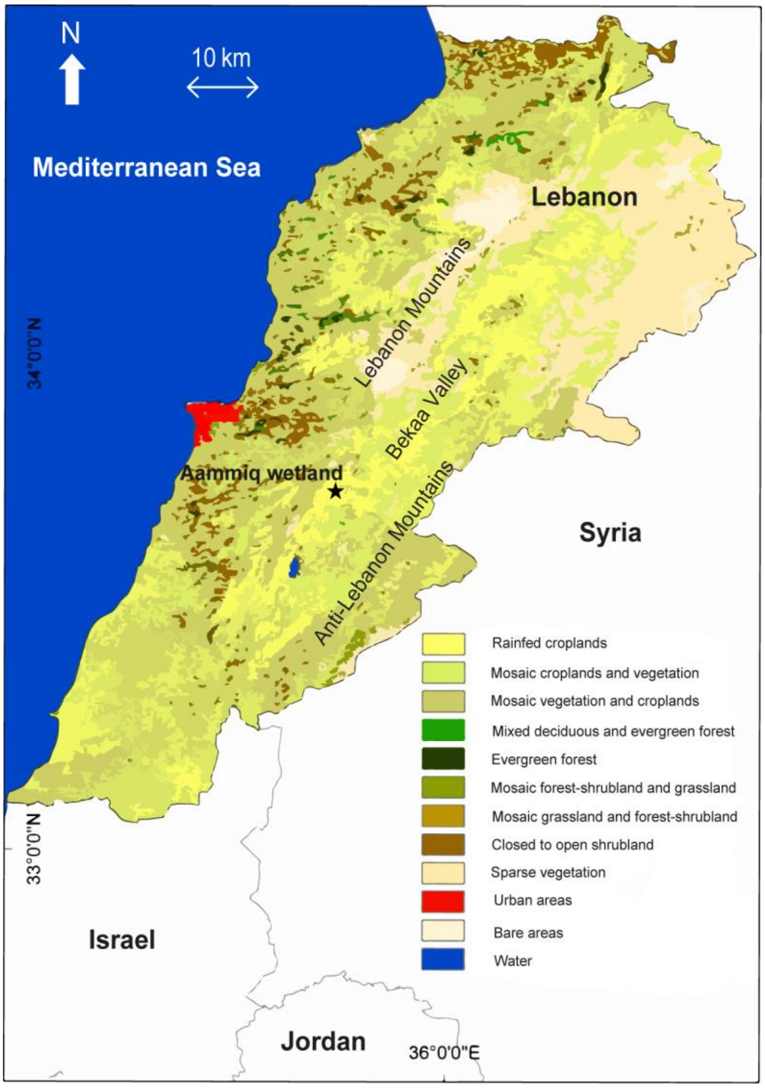

Figure 7.1 Land cover map of Lebanon constructed from GLOBCOVER information (Bicheron et al., 2009) showing location of Aammiq Wetland.

The climate is essentially Mediterranean in character with warm dry summers and cool wet winters. However, the climate is semi-arid due to the rain shadow effect of the Lebanon Mountains (Sene et al., 1999) that relieves the westerly rain bearing extratropical cyclones, that are the main source of precipitation (Shay-el and Alpert, 1991) of most of their moisture. As a consequence, annual mean precipitation is approximately 600 mm yr^{-1} with inter annual variability of 30% to 200% from the average (Storey, 2007). On the Mountains annual precipitation can reach 1600 mm yr^{-1} on the upper ridge with volumes declining on the eastern slopes with decreasing altitude (Asmar, 2011). A majority of the precipitation falls from January to April in the valley as rainfall and on the mountain slopes as snow. Average temperature in the valley during the coldest months (December, January, February) is 2 °C and during the warmest (July) 33 °C. On the higher mountains slopes temperatures are below 0 °C in winter and can exceed 30 °C in summer (Gasse et al., 2011).

The contemporary wetland is ephemeral and can encompass up to 300 ha during periods of heavy rainfall, but only approximately 70 ha remains flooded continuously during the four month wet season (Storey, 2007). From late July to December with the exception of a man-made deep pool the wetland is usually dry (Storey, 2007). The wetland is a mosaic of different habitats including reed beds comprised predominately of *Phragmites australis* (common reed), *Typha spp*. (reed-mace) and *Scirpus* (rushes) and open water pools containing numerous sub-merged, emergent and floating aquatic herbs (Conroy and Hepper, 2005). Surrounding the wetland on the Bekaa plain are agriculture fields. The vegetation on the eastern slopes of the Lebanon Mountains can be differentiated by altitude into three vegetation belts. On the lower slopes from 800 m to 1000 m a.s.l (Eu-Mediterranean belt) an overgrazed rangeland is present, comprised of garrigue (scrubland) and exposed sub-desertic soils (Asmar, 2011). From 1000 m to 1500 m (Supra-Mediterranean) is predominately comprised of heavily grazed evergreen *Quercus calliprinos* (Palestine oak) and *Quercus infectoria* (Aleppo oak) forests (Asmar, 2011). Above 1500 m (Mountainous-Mediterranean) sparse *Juniperus excelsa* (Greek juniper) trees exist amongst thorny shrubs (Asmar, 2011).

7.4 Methods

A 7 m sediment core was extracted from the wetland (33°43'51.47"N, 35°47'08.80"E) using a Russian corer in November 2008. To reduce moisture loss prior to laboratory analysis the core was wrapped in cellophane and aluminium foil and black plastic tubing to reduce exposure to sunlight so that Optically Stimulated Luminescence (OSL) dating could be conducted. Pollen analysis (see Jeffers and Willis, submitted) has previously been conducted on this core from which the charcoal data for this study was obtained.

Charcoal analyses

For macrocharcoal analysis, 1 cm^3 of sediment were taken contiguously at 1 cm resolution from the sedimentary core, yielding a total of 206 samples. Samples were disaggregated in sodium hydroxide (NaOH) and isolated by washing through a 150 µm sieve. The residue was transferred into a petri dish with the base divided into a grid and the charcoal particles were counted using a stereo microscope at × 40 magnification. Microcharcoal particles < 150 µm were counted on the slides prepared for pollen analysis. Microcharcoal analysis was conducted following the method proposed by Tinner and Hu (2003); with charcoal particles along with the exotic added pollen grains (*Lycopodium*) counted until their combined sum totalled 200.

Dating

The chronology of the core was established by a combination of optically stimulated luminescence dating (OSL) conducted at the Oxford Luminescence Laboratory, University of Oxford, UK and radiocarbon dating at The ^{14}CHRONO Centre, Queen’s University Belfast, UK. Luminescence dating was conducted at five depths: 219, 278, 486, 544 and 690 cm. In the laboratory, sediment (1 cm^3) was extracted from the centre of the sediment core under subdued yellow laboratory lights. Quartz grains ranging from 75-210 µm in size were extruded from the sediment (219, 278, 544 and 690 cm) using standard methods (Telfer and

Thomas, 2007, Thrasher et al., 2009) which initially included 30% HCl and 30% H_2O_2 to remove carbonates and organics respectively, followed by 60% HF to remove feldspars and the outer layer of the quartz grain affected by alpha radiation. The quartz was then wet sieved to isolate grains 75-210 µm in size. Fine grain quartz (5-15 µm) was retrieved from the sediment sample from 486 cm by applying additional wet sieving and density separation through flotation in H_2O to the remaining quartz fraction. Grains were analysed using the Single Aliquot Regenerative (SAR) protocol of Murray and Wintle (2000, 2003). Measurements were made on a Risø TL-DA-15 reader using blue diode illumination and U-340 filters.

Radiocarbon dating was performed on bulk sediments from two depths: 258 and 360 cm that were sent to ^{14}CHRONO for sample preparation and Accelerator Mass Spectrometric analysis (AMS).

Development of an age model

Calibration of the bulk sediment radiocarbon dates (^{14}C ages) to years before present (cal. yr BP) was performed using the software CLAM version 2.1 (Blauuw, 2010) default northern hemisphere (atmospheric) calibration curve that uses IntCal09 (Reimer et al., 2009). The Age-depth model for the sedimentary sequence combining the calibrated radiocarbon dates and the OSL dates was then constructed using linear interpolation in CLAM run in R programming language (2010).

Numerical analysis

Spectral analysis of the time series data was conducted using the software PAlaeontological STatistics (PAST) version 2.17 (Hammer et al., 2001) that includes a Lomb periodgram algorithm to identify periodicities within the macro and microcharcoal data. The advantage of using this technique over a Fourier-base method is its applicability to unevenly spaced data (Press et al., 1992). Cross correlation was also performed on these data to evaluate temporal relationships between the charcoal and pollen time series data, and to evaluate whether time lags or leads between the correlations exists (Green, 1981) .

7.5 Results

Chronology

Results of the age-depth model combining radiocarbon and optically stimulated luminescence dates are presented in Table 7.1 and 7.2 respectively and Figure 7.2. The time period covered by this study is ~ 104 to ~ 35 ka BP corresponding to the section of sediment core 544-338 cm.

Sample depth (cm)	Laboratory number	Sample Material	^{14}C age yr. BP	2σ Calibrated age (yr BP)
258	UBA-20316	Bulk sediment (Clay/organic rich)	12836 ± 69	14939-15882
360	UBA-20315	Bulk sediment (Clay/organic rich)	37205 ± 745	41058-43108

Table 7.1 Radiocarbon dating results from the Aammiq Wetland sedimentary sequence. ^{14}C ages were calibrated using IntCal09 (Reimer et al., 2009).

Sample depth (cm)	Lab code	Grains accepted to calculate D_e	Minimum Maximum grain size (µm)	K (%)	U (ppm)	Th (ppm)	Cosmic dose rate (Gy/ka)	Total dose Rate (Gy/ka)	D_e (Gy)	Age (ka)
690	OxL 2270	16	75-125	1.04 ± 0.05	4.06 ± 0.46	15.69 ± 1.56	0.10 ± 0.01	2.87 ± 0.18	355.56 ± 30.03	124 ± 13.2
544	OxL 2268	18	75-125	0.27 ± 0.007	1.91 ± 0.08	3.92 ± 0.25	0.12 ± 0.01	1.01 ± 0.05	105.6 ± 8.25	104.6 ± 9.5
486	OxL2267	12 *	5-15	0.38 ± 0.02	3.4 ± 0.2	5.2 ± 0.26	0.10 ± 0.01	1.86 ± 0.17	163.0 ± 5.57	87.8 ± 8.6
278	OxL2271	16	75-125	0.25 ± 0.12	3.4 ± 0.34	1.0 ± 0.1	0.16± 0.01	0.81 ± 0.05	16.28 ± 1.02	20.1 ± 1.7
219	OxL 2263	24	75-125	0.48 ± 0.02	4.52 ± 0.45	10.32 ± 1.0	0.18 ± 0.01	2.21 ± 0.15	11.38 ± 0.48	5.1± 0.4

Table 7.2 Results of optically stimulated luminescence dates. * D_e (equivalent dose) calculated from 12 aliquots of fine grain quartz, assumes a moisture content of 15 ± 5%. Moisture content of all other samples assumed to be 10 ± 3 %.

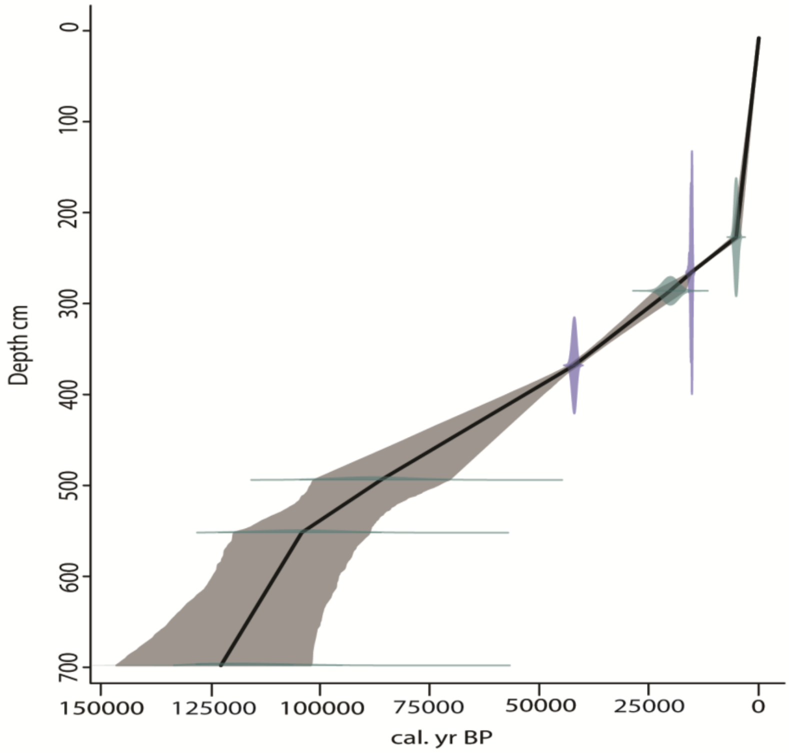

Figure 7.2 Age-depth graph undertaken in CLAM version 2.1 (Blaauw, 2010) using linear interpolation. Black line represents the 'Best fit' model and the grey shaded area 95% confidence limits. The probability distribution of the calibrated radiocarbon dates are shown as blue reduced plots, and the probability distribution of the OSl dates as green reduced plots.

Macro and microcharcoal

Analysis of the macrofossil charcoal particle concentrations (particles cm^{-3}) demonstrates considerable variation in concentrations through time ranging from 0 to ~ 13876 particles cm^{-3} (Fig. 7.3). Significant peaks occur between 473 cm and 465 cm (~ 82 to ~ 78 ka BP) and 457 cm and 454 cm (~ 76 to ~ 75 ka BP) with lesser but still significant maxima at 505 cm (~ 92 ka BP) and from 402 to 353 cm (~ 58 to ~ 39 ka BP). No macrocharcoal deposition is recorded periodically throughout the sediment record with extended periods between 348 and 338 cm (~ 38 and ~ 35 ka BP), 504 to 500 cm (~ 92 to ~ 92 ka BP) and 510 to 515 cm (~ 95 to ~ 93 ka BP). Spectral analysis of this time series identified a cyclical behavior in the time series data at or above the 95% confidence interval as occurring every 40 ka and 21 ka (Fig. 7.4). No significant positive correlation was identified between macrocharcoal and any individual plant taxa or aggregate of steppic herbs, non-steppic herbs or for the sum of arboreal pollen.

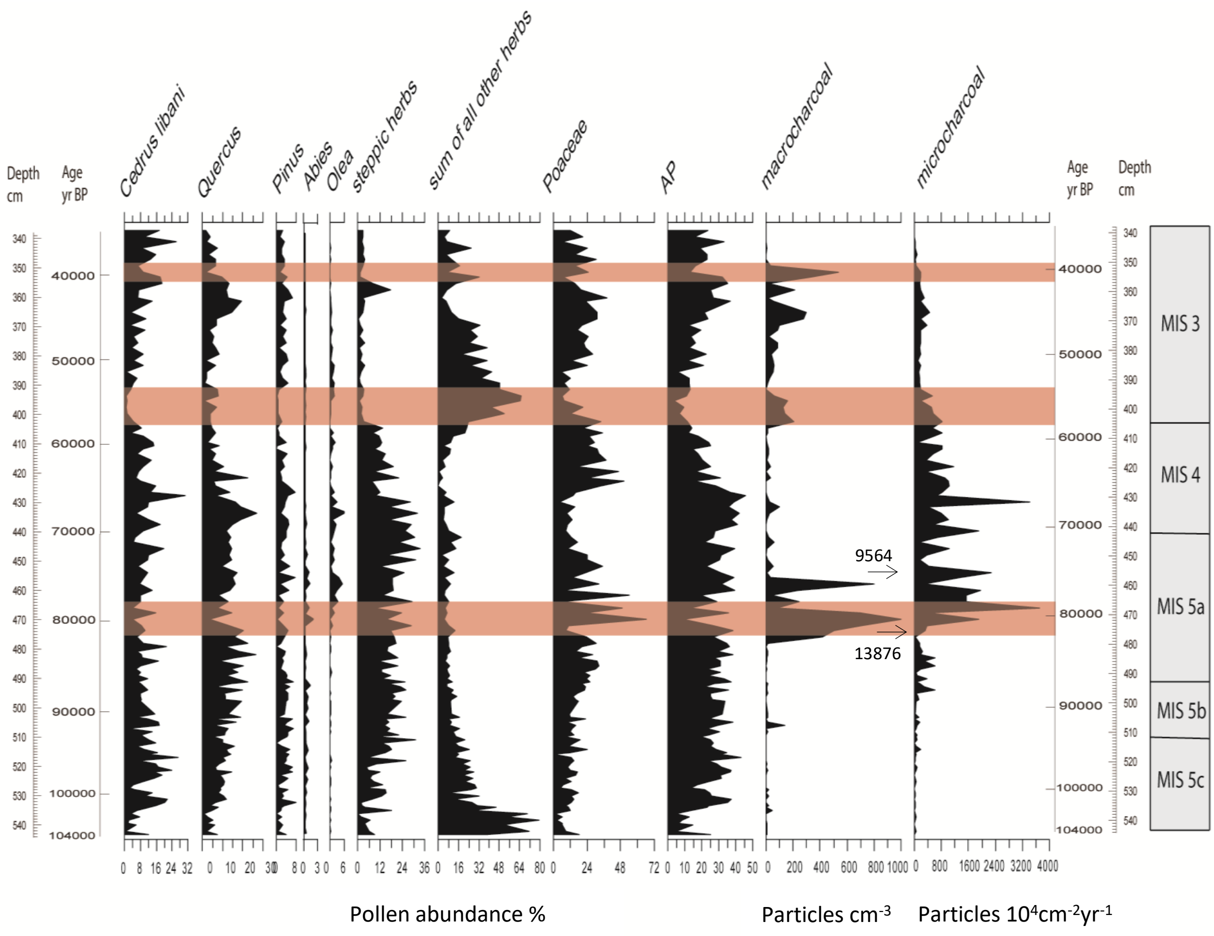

Figure 7.3 Macro and microcharcoal influx data for Aammiq Wetland plotted against age and depth and approximate marine isotope stage (MIS) boundaries using C2 version 1.7.2 (Juggins, 2011) are compared to a summary of the pollen data (from Jeffers and Willis, submitted). The horizontal red shaded bars indicate periods in time (~ 82 to ~ 78 ka BP, ~ 58 to ~ 53 ka BP and ~ 40 to ~ 39 ka BP) when peaks in fire activity correspond to periods of precession minima.

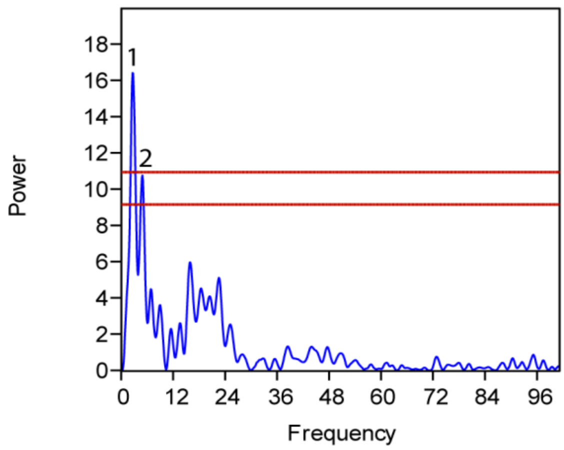

Figure 7.4 Spectral analysis performed in PAST version 2.17 (Hammer et al., 2011) on the macrocharcoal time series data (~ 104 to ~ 35 ka BP). Red horizontal lines represent 0.05 and 0.01 confidence intervals. The numbered peaks on the frequency axis (X-axis) equate to frequencies of (1) 2.5 and (2) 4.8 and the power axis (Y-axis) units are proportional to the square of the amplitudes of the sinusoids present in the data. Given that the frequency axis is in units of 1/(x units) the periodicities in fire return intervals therefore equate to (1) 40 ka and (2) 21 ka.

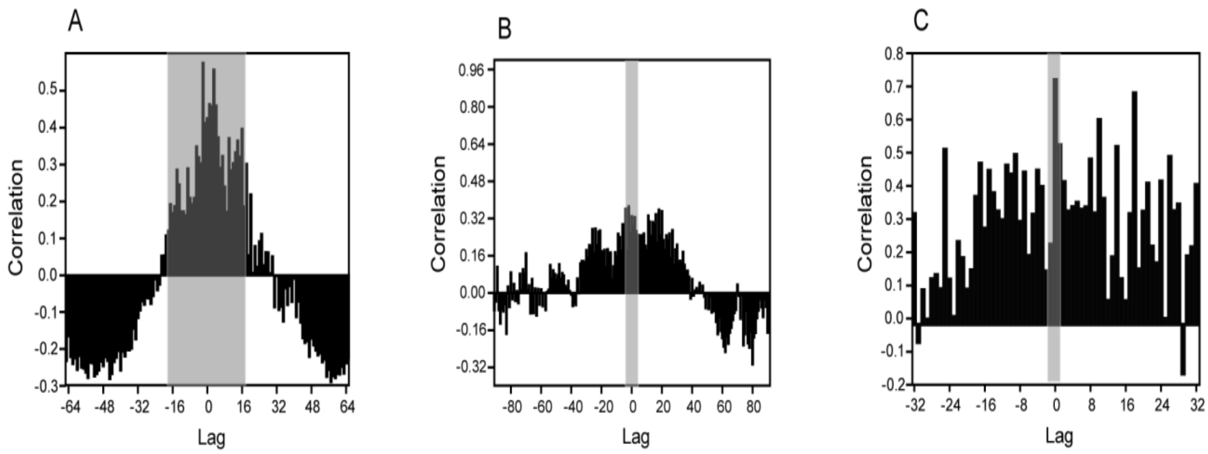

Figure 7.5 Cross-correlograms constructed in C2 version 1.7.2 (Juggins, 2011).Vertical shaded bars show when correlation was significant at *P*=0.01. (A) microcharcoal and *Olea* (r=>+0.42) and (B) microcharcoal and Poaceae (r=0.33) for the entire time series (~ 104 to ~ 35 ka BP) and (C) microcharcoal and Poaceae (r=0.73) for the time period ~ 104 to ~ 82 ka BP.

Microcharcoal accumulation rates CHAR (particles cm^{-2} yr^{-1}) are displayed in Fig. 7.3. Very low levels of microfossil charcoal influx are recorded during the last glacial period before ~ 90 ka BP and after ~ 37 ka BP. An increase in microcharcoal accumulation rates occurs after ~ 90 ka BP (497 cm) with a significant increase after ~ 82 ka BP (471 cm) with a peak in CHAR of 3708 particles cm^{-2} $yr^{-1}10^4$ obtained at 467 cm (~ 79 ka BP). Between ~ 82 and ~ 45 ka BP (404 cm) there are 14 observable peaks in microcharcoal. A significant positive correlation for the entire sequence exists between microscopic charcoal and *Olea* (r=>+0.42 at 0 lag to +4 lags and 0 to -2 lags) and Poaceae (r=0.33, p=0.01 at 0 lags) and between microcharcoal and Poaceae for the period ~ 104 to ~ 82 ka BP (544 to 470 cm) r=0.73, p=0.01 at 0 lags (Fig.7.5). Applying spectral analysis to the microcharcoal time series data did not identify any periodicity in the peaks of charcoal influx.

7.6 Discussion

Results from this study indicate that there are a number of trends discernible in the fire record from this region during the interval between ~ 104 to ~ 35 ka BP and these appear to have been linked to specific abiotic and biotic drivers.

High tree biomass and low fire activity

Local and regional fire activity (Fig. 7.3) was infrequent and of apparently low intensity during the early glacial period marine isotope stage (MIS) 5c to MIS 5b (~ 104 to ~ 90 ka BP). Previous palaeoclimatic studies from Yammoûneh Basin, northern Lebanon have shown that wet climatic conditions prevailed during this period (Develle et al., 2011, Gasse et al., 2011). The wet conditions favoured the growth of an oak and coniferous forest comprised of *Cedrus libani*, *Pinus* and *Abies* on the slopes of the Lebanon Mountains (Jeffers and Willis, submitted). The presence of *Abies* suggests that precipitation levels likely exceeded 1000 mm yr^{-1} (Aussenac, 2002). Increased levels of woody biomass that occur during wet periods in the eastern Mediterranean have previously been associated with an increase in fire frequency and intensity (Turner et al., 2010) as a result of an increase in fuel loads. Our result finds the opposite effect and demonstrates that during high precipitation periods an increase in woody biomass appears to have led to low frequency and intensity fires. We propose that in this region increased precipitation dampened fire activity. This effect of high precipitation on tree fire interactions is possibly mediated through the accumulated leaf litter under the tree canopy. During relatively wet periods (as in MIS 5c and MIS 5b), it has previously been suggested that un-decomposed leaf litter in the understory has a higher moisture content (Chang, 1996) and this acts as a fire barrier, thus reducing both the likelihood of fire ignition around trees and the intensity of burning when fire occurs. The tree canopy further acts to maintain a moist leaf layer in the understory by reducing sunlight penetration and air circulation that would otherwise dry out the forest litter fuels (Cochrane et al., 1999). This has also been observed in the southern Mediterranean, where the expansion of forests (Rossignol-Strick, 1999, Tinner et al., 2009, Magny et al., 2012) had a positive association with low fire activity when humid conditions prevailed during the mid-Holocene ‘Thermal Maximum’ (Vanniere et al., 2011).

Grass-fire feedback mechanism

High levels of grass biomass are typically associated with frequent but low intensity fires (Naveh, 1975, Vilà et al., 2001, Bond and Keeley, 2005). Here we found a significant, positive correlation between grass pollen abundance and microcharcoal ($r=0.73$, $p=0.01$ at 0 lags) for the time period (~ 104 to ~ 82 ka BP). This is particularly evident during the period ~ 90 to ~ 82 ka BP (Fig. 7.3) when peaks in regional fire activity were concurrent with an increase in the proportion of grass. A positive feedback mechanism therefore appears to exist during this period in time between fire and grass, where the almost continuous presence of flammable grass litter maintains a fire regime of low intensity fires that occurs regularly, particularly during periods of vegetation senescence. Previous studies have suggested that the presence of a fire-grass feedback mechanism is facilitated by a number of factors including grass seed germination being stimulated by moderate heat (Reyes and Trabaud, 2009) and exposure to smoke (Staden et al., 2000), its fast vegetative growth in post fire nutrient-rich soils (Vilà et al., 2001) and for most annual and perennial species the annual cycle of die back and regeneration. This feedback mechanism has been shown by Vilá et al. (2001) in NE Spain to have facilitated the expansion of grass when fires were infrequent and of low intensity; we propose a similar process was occurring in the Aammiq region between ~ 90 to ~ 82 ka BP.

Orbital forcing of fire activity

The peak local and regional fire activity during the last glacial period occurred during periods of precession minima. The highest values of macrocharcoal concentrations (13876 cm^{-3}) and microcharcoal accumulation rates (3708×10^4 particles cm^{-2} yr^{-1}) were recorded during the early glacial period between ~ 82 and ~ 78 ka BP. The timing of this increase in both local and regional biomass burning corresponds to a period (~ 84 to ~ 78 ka BP) of precession minima (Fig. 7.3), when maximum insolation in the northern hemisphere occurred during the summer (Tuenter et al., 2005). Previous work has demonstrated that at the start of precession minima periods, the Mediterranean Basin experienced regional aridity (Tzedakis, 2007). At Aammiq Wetland, this arid period ~ 82 and ~ 78 ka BP was associated with a shift to steppic herbs and grasses and a reduction in *Quercus*. Increased levels of dry herbaceous plant biomass are typically associated with an increase in the frequency (Naveh, 1975, Vilà et al., 2001) and spatial extent of fires (Brooks et al., 2004). This summer aridity would also have resulted in a decrease in the moisture content of the canopy foliage (Agee et al., 2002), and as a consequence an increase in the flammability of the trees. Previous studies have demonstrated that greater flammability during the summer leads to high fire intensity and high rates of tree mortality (Belsky and Blumenthal, 1997, Duffin, 2008). The observed reduction in oak may therefore be a consequence of increased frequency and intensity of the fire regime at this site. Although Mediterranean oaks are extremely resilient to fire due to a strong resprouting capacity (Rodrigo et al., 2004, Pausas et al., 2008) increasing fire recurrence is known to result in reduced oak population density (Pausas and Vallejo, 1999).

The reduction in oak tree density likely facilitated the expansion of (wild) olive (*Olea europaea* L.var. *sylvestris*) at ~ 82 ka BP. Olive can only coexist with oak only in an open landscape (Lumaret et al., 2004) because olive is light demanding (Rhizopoulou et al., 1991); reduced oak tree density significantly increases light penetration to understorey plants (Tyson, 1995). A reduction in canopy cover may have created a feedback mechanism that maintains openness by increasing the susceptibility of the remaining trees to fire (Cochrane et al., 1999). This occurs when greater amounts of insolation can reach the leaf litter layer, thus resulting in drier and more flammable biomass (Pellizzaro et al., 2007). The presence of a highly flammable litter layer around trees can then act as a vector for the transfer of surface to crown fires (Agee et al., 2002), increasing tree mortality rates and maintaining an more open landscape. This new regime of high intensity and frequency fires appears to have persisted for approximately 3000 years, concurrent with the duration of this period of precession minima.

This orbital forcing of fire activity was a recurrent event as indicated by spectral analysis of the macrocharcoal concentrations (Fig. 7.4). Our results show that peaks in local fire activity for the duration of the last glacial period occurred at a frequency of ~ 21 ka and ~ 40 ka, corresponding with the return period (and double return period) of precession minima in the northern hemisphere that occurs with a periodicity of 21 ka (Lowe and Walker, 1997, Bradley, 1999). The first occurrence spanned ~ 82 to ~ 78 ka BP (as described above) and was followed by two other periods of peak fire activity between ~ 58 to ~ 53 ka BP and ~ 40 to ~ 39 ka BP. As before, these peak fire periods were concurrent with the expansion of herbs (~ 58 to ~ 53 ka BP), and herbs and grasses (~ 40 to ~ 39 ka BP) and a reduction in tree cover associated with both of these intervals in time. As with the previous period of precession minima during these events maximum insolation occurred during the summer. However, in contrast, the latter two peaks in fire activity (~ 58 to ~ 53 ka and ~ 40 to ~ 39 ka BP) occurred during a period in time when previous work has suggested that temperatures were 3 to 5 ºC lower than present day (Ayalon et al., 2013) and moisture availability had declined significantly (Develle et al., 2011, Gasse et al., 2011) during the full glacial period. We propose that at Aammiq this change in climatic parameters relative to the first period of precession minima (~ 84 to ~ 78 ka BP) resulted in lower levels of vegetation productivity and thus lower levels of herb and grass biomass. The presence of lower herbaceous biomass between ~ 58 to ~ 53 and ~ 40 to ~ 39 ka BP would have created an accumulation of highly flammable and rapidly burnt material (Cofer et al., 1990), that lacked the fuel load and therefore the intensity of fire activity recorded during the earlier period of precession minima.

Millennial-scale periodicity in peak regional fire activity

Within these orbital forced periods of climate change, our results indicate that there existed shorter-term, millennial-scale peaks in regional fire activity between ~ 75 and ~ 45 ka BP which possibly correspond in timing and duration to Dansgaard-Oeschger (D-O) climatic warming events 20 to 14 (Fig. 7.6), recorded in the Greenland ice-cores (Dansgaard, 1993) and north Atlantic marine cores (Bond and Lotti, 1995) as ice rafted debris events. During these D-O events temperature increases in the North Atlantic region exceeded 6 ºC, which is more than half of the temperature increase recorded during an interglacial-glacial cycle (Sakai and Peltier, 1999). As a consequence of the presence of a particularly strong teleconnection between the North Atlantic region and the Mediterranean Basin (Eshel and Farrel, 2000) the presence of high pressure over Greenland is consistent with anomalous EM meridional southerly winds that advect warm air into the EM (Eshel and Farrel, 2000). Warmer climatic conditions in the EM during D-O interstadials would be expected to increase the tendency for higher intensity fire activity due to the prevalence of dry climatic conditions (Vázquez and Moreno, 1993, de Luís et al., 2001).

The peaks in fire activity are also concurrent with changes in vegetation abundance, which indicates a closely coupled relationship between vegetation, fire and D-O events. During D-O events and peaks in fire activity between ~ 75 and ~ 45 ka BP, grass abundance

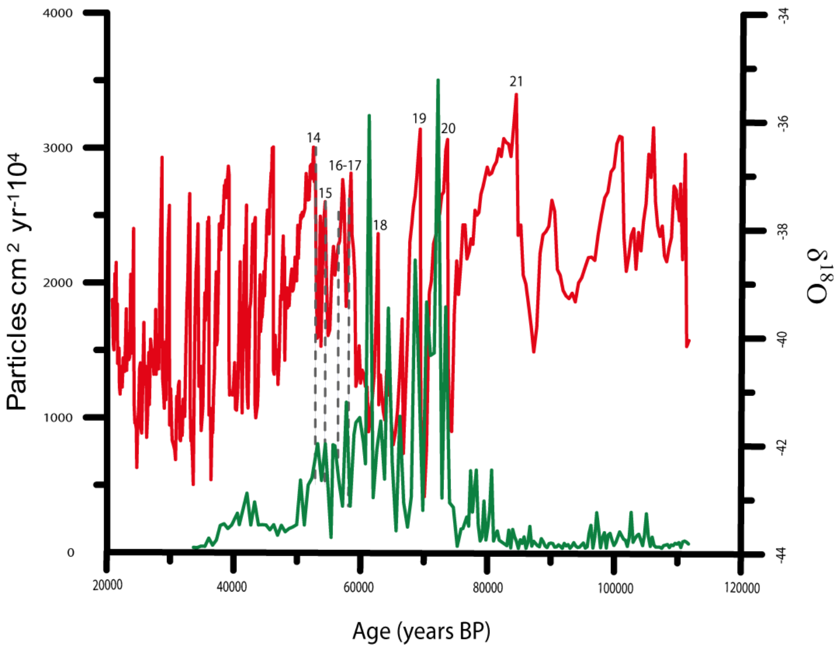

Figure. 7.6 Graph showing last glacial microcharcoal data (~ 104 to ~ 35 ka BP) from Aammiq Wetland (green line) compared with Greenland Ice Sheet Project (GISP) oxygen isotope data (red line) (Grootes et al., 1993 and 1997). Note the apparent concurrence of the peaks in microcharcoal influx and Dansgaard-Oeschger (D-O) events 20-14.

declined while Mediterranean forest taxa (oaks and olive) increased in abundance. Expansion of trees during D-O events has also been observed in southwestern Iberia (Daniau et al., 2007). This phenomena has been attributed to warming-induced increases in forest productivity (Daniau et al., 2010), and that higher biomass fuelled more intense fires (Daniau et al., 2007). At Aammiq D-O events and peak fire were associated with greater abundance of olive (Fig. 7.5), although wild olive trees are highly inflammable (Spennemann, 1998) their increased abundance increases fire frequency (Parsons and Cuthbertson, 1992) due to the high proportion of dead and dry fruitwood within the canopy (Spennemann, 1998).

Cold temperatures and low fire activity

The incidence of local fire activity ceased at ~ 38 ka BP and regional fire activity declined significantly after ~ 37 ka BP. This finding is consistent with the observed decrease in global biomass burning at the end of the last glacial period during the last glacial maximum (Power et al., 2008). The decline in fire activity corresponds with a reduction in temperatures (during MIS 3 - MIS 2) in the Lebanon Mountains that were 6-10 °C lower than present day (Develle et al., 2010), which would have resulted in average summer temperatures of approximately 20 ºC in the Bekaa Valley and 10 ºC in Lebanon Mountains. This decline in temperatures would have been associated with declining evaporative losses, which would have reduced plant moisture loss (Hopkins and Huner, 2008). The effect on fire activity would be a decline in biomass ignitability and flammability (Dimitrakopoulos, 2002), and thus reduced fire frequency and intensity at the local and regional spatial-scales during the later stages of MIS 3.

7.7 Conclusions

This new study of fire history provides an important record of local and regional fire activity in the Bekaa Valley and the mountainous regions of Lebanon during the last glacial period. The major findings of this study are as follows.

Low local and regional fire activity during the early glacial period (~ 104 to ~ 90 ka BP) shows a strong association with the wet climatic conditions. This climate regime favoured the growth of oak and coniferous forests, with high moisture content and damp leaf litter layers that acted as a fire barrier. In contrast the increase in regional fire activity between ~ 90 and ~ 83 ka BP shows a significant correlation with an increase in abundance of grass. That provided a highly flammable and annually renewable fuel source that resulted in a high frequency and low intensity regional fire regime.

The micro and macocharcoal records that attain their highest values between ~ 82 and ~ 78 ka BP show a close relationship with the precession minima associated summer insolation increase. The resulting increase in aridity facilitated the expansion of steppic herbs and grasses that provided an abundant and rapidly renewed source of highly flammable dry plant biomass. The concurrent observed reduction in oak and opening of the canopy may be a consequence of the increase in the frequency and intensity of the fire regime. This reduction in canopy cover created a feedback mechanism that increased the susceptibility of the remaining trees and leaf litter layer to fire. The recurrence of periods of peak local fire activity at ~ 58 to ~ 53 ka BP and ~ 40 to ~ 39 ka BP also occurred during periods of precession minima. In contrast with the peak in fire activity at ~ 82 and ~ 78 ka BP the later peaks in fire activity occurred against a background of lower temperatures and moisture availability as the glacial period progressed towards the last glacial maximum (LGM). This climate regime increased the frequency and intensity of fire, but at a lower intensity relative to the earlier period of precession minima.

The millennial-scale peaks in regional fire activity between ~ 75 and ~ 45 ka BP correspond closely in timing with D-O warming events recorded in the Greenland ice cores. The peaks in regional fire activity are concurrent with a decrease in grass abundance and an increase in Mediterranean tree taxa suggesting that during D-O events woody biomass accounted for a significant amount of the fuel load.

The cessation of local fire activity at ~ 38 ka BP and significant decline in regional fire activity at ~ 37 ka BP reflects the reduction in insolation levels and associated decline in regional temperatures during the later stages of MIS 3.

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Chapter 8: Conclusions and synthesis of palaoenvironmental data from Aammiq Wetland

This chapter provides a summary and synthesis of the findings of the four papers that together with the introduction, study site and methods form this thesis. The research undertaken for this thesis is the first to evaluate the climate and environmental dynamics of the southern Bekaa Valley and southern Lebanon Mountains from the end of the last interglacial and through the last glacial period (~ 112 to ~ 35 ka BP). This has been achieved using a wide range of palynolological and palaeoenvironmental techniques and numerical analyses. The thesis aimed to address four key knowledge gaps relating to long-term climate, vegetation and landscape dynamics of this important area in the eastern Mediterranean and the findings are as follows:

How did the climatic conditions of the last glacial period vary in southern Lebanon? What was the effect of this climatic variability on landscape processes and the wetland in the Bekaa Valley?

At Aammiq, the geochemical composition, mineral magnetic properties, loss-on-ignition values of the sediment and changes in sediment type provided an important record of the climatic and environmental events which characterised the end of the last interglacial and the last glacial period in this mountainous lake catchment. At the end of the last interglacial and the start of the last glacial period (~ 112 to ~ 85 ka BP), there was a high level of water circulation through the karstic system into the lake. This was predominately a time of high forest cover and stable soils with associated low levels of landscape erosion. This period of long-term stability of the watershed’s surface sediments was interrupted by periods of high erosion. The first of these that occurred between ~ 108 to ~ 104 ka BP was associated with high quantities of wind-blown clastic material into the lake. The second from ~ 84 ka BP led to the development of an anaerobic wetland/swamp that received greater input of fluvial derived sediments. During the full glacial period ~ 70 ka BP there was a major shift within the catchment from high soil stability and vegetation cover to increased erosion, reduced vegetation cover (as identified in **Chapter 6)**, and multiple changes in sediment type. During the later periods of the glacial period there was an increase in local water availability. That is assumed to derive from an increase in precipitation and thus water circulation through the karstic system into the wetland.

What was the plant community response to glacial conditions after the last interglacial period (~ 112 to ~ 104 ka BP)?

Analyses of pollen, isotopes of carbon and total organic carbon show a significant shift in the composition of the vegetation community during the transition period from the last interglacial to the last glacial period. At the end of the last interglacial ~112 to ~ 110 ka BP a mixed coniferous-deciduous forest-grassland dominated by *Cedrus libani* covered the slopes of the Lebanon Mountains. The persistence of this forest was facilitated by high levels of precipitation as described in **Chapter 4.** The presence of temperate deciduous species (*Betula*, *Alnus* and *Corylus*) in the forest matrix also indicates that there was not a strong summer seasonal water deficit as exists during the current interglacial.

At the start of the glacial period (MIS 5d, ~ 110 ka BP) an environmental threshold was crossed in response to a significant reduction in precipitation, as identified in **Chapter 4.** The arid conditions lead to the loss of the temperate deciduous arboreal species (*Betula*, *Alnus* and C*orylus*) and significant reduction in the populations of coniferous trees *Cedrus libani* and *Pinus*. The considerable decline in water availability favoured the development of grassland/steppe dominated by Chenopodiaceae and *Artemisia*.

How did climatic variability throughout the last glacial period alter the structure and composition of plant communities in the Bekaa Valley and on the Lebanon Mountains?

The composition of the vegetation community on the Lebanon Mountains and in the southern Bekaa Valley varied significantly during the last glacial period in response to climatic variability. During MIS 5c (~ 102 ka BP) in response to an increase in precipitation the herbaceous steppe was abruptly replaced by a forest/grassland, that persisted up to ~ 82 ka BP. This vegetation mosaic was subsequently lost during the period of precession minima at ~ 82 to ~ 78 ka BP when oak trees were lost and grass expanded. Two hypotheses were proposed to explain the decline in forest: 1) the loss of trees may have been driven by the decline in water availability which was indicated by the decline in water levels within the lake and the subsequent development of a wetland/swamp as argued in **Chapter 4;** 2) or as suggested in **Chapter 7** was a response to an increase in the frequency and intensity of fire activity which itself was driven by the increase in aridity.

Between MIS 5a and MIS 3 (~ 74 to ~ 40 ka BP) the expansion of trees appears to occur in millennial-scale cycles that possibly corresponds in timing to Dansgaard-Oeschger (D-O) warming events. As described in **Chapter 7** this may be the result of warming induced increases in forest productivity within a trend of declining temperatures as the last glacial period advanced as identified in **Chapter 4**. Interestingly the expansion of forests on the southern Lebanon Mountains to its maximum extent during the glacial period at ~ 67 ka BP did not occur during a D-O event. This period of 'forest maxima' was followed by the contraction of forest, so that by ~ 54 ka BP the landscape in the Aammiq region was almost totally devoid of trees. During the later stages of MIS 3 as detailed in **Chapter 4** there was a slight increase in effective precipitation that was sufficient to enable the development of a grassland-forest. The presence of *Cedrus libani* pollen throughout this sedimentary sequence identifies the importance of southern Lebanon as a climatic refuge during the last full glacial period for this emblematic tree species.

What was the frequency of fire activity in and around Aammiq Wetalnd during the last glacial period? How did this relate to climatic variability? And what was the impact of these disturbances on the plant community?

Variations in macro and microcharcoal abundances in the sediments from Aammiq Wetland were used to reconstruct the local and regional fire regimes respectively. These data indicate that during the last glacial period fire activity was intrinsically linked to the prevailing climate regime. Between (~ 104 to ~ 92 ka BP) the frequency and intensity of local and regional fire activity was constrained by high levels of moisture availability. The presence of forest as described in **Chapter 6** and high levels of precipitation as shown in **Chapter 4** resulted in the presence of living biomass with a high water content and accumulation of moist leaf litter under the tree canopy that acted as a fire barrier. This would have reduced both the likelihood of fire ignition around trees and the intensity of burning when fire occurred.

The significant increase in local and regional fire activity between ~ 82 and ~ 78 ka BP was concurrent with the decline in precipitation as identified in **Chapter 4** and expansion of grasses as described in **Chapter 6**, that were associated with a period of precession minima. This process was repeated every ~ 21 ka or so, during subsequent periods of precession minima. Within these long-term patterns, there were also millennial-scale variations in fire activity which were coincident with D-O warming events 20-14. At this temporal scale increased fire activity appears to lead to more oak and less grass cover.

Future work

The work presented in this thesis provided the longest and most accurately dated vegetation record in southern Lebanon spanning ~ 112 to ~ 35 ka BP. It showed that the iconic plant communities on the Lebanon Mountains have persisted for the duration of the last glacial period, despite many long-term shifts in community composition. My results showed a tight coupling between these changes and relative shifts in aridity and fire activity. Yet, there is still more to learn about the vegetation community in this semi-arid environment.

How far back in time might the sedimentary record in the Bekaa Valley extend?

The sedimentary sequence that provided the data for this thesis when extracted was 7 m in length and covered the end of the last interglacial and a majority of the last glacial period. To put this in context it is believed that the depth of sediments in the Bekaa Valley may exceed 1000 m. The extraction and analysis of these sediments would therefore provide a record of climatic and environmental change in the EM that covered multiple interglacial-glacial cycles, perhaps even the entire Quaternary period.

How did the spatial extent of the Aammiq Wetland vary through time?

The evidence of the decline in water input into the wetland provided in **Chapter 4** indicates that the wetland expanded and contracted in rhythm with changes in climate. It would therefore be helpful to collect a number of sediment cores from across the Bekaa Valley to determine how the spatial extent of the wetland varied through time under varying climatic conditions.

Improving age control

Two factors have made this research challenging: achieving an accurate chronology and extracting enough identifiable pollen grains from the sediment. Optically stimulated luminescence provided a reliable chronology for ecosystem dynamics at Aammiq that were consistent with regional patterns. However, this chronology could be further constrained with analysis of tephra deposits in the sediments (if they are present) and their assignment to known and well dated volcanic events in the central and eastern Mediterranean Basin

Final comments

Given expected future scenarios of climate change in the EM as forecasted by the IPCC (2007), this region should experience warmer and drier conditions than in the present time. My 80,000 year long record showed that in the past, these conditions have resulted in major changes in plant community composition. However, the plant communities shifted continuously between the same groups of taxa i.e grasses and herbs in dry periods and coniferous-deciduous forest in wet periods. Thus, in the EM vegetation communities appear to have a high level of resilience to extreme shifts in climatic conditions. Although major vegetation changes are likely to occur, as long as natural processes of regeneration are allowed to proceed, these will once again be reversed once anthropogenic climatic warming has passed.