Thirlmere Lakes; A degraded environment or an environment sensitive to natural hydroclimate variability?

J Allenby

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Thirlmere Lakes; A degraded environment or an environment sensitive to natural hydroclimate variability?

Abstract
The Thirlmere Lakes are a series of five freshwater lakes trapped in an ancient uplifted river meander, located on the eastern margin of the Greater Blue Mountains World Heritage Area, near Thirlmere, NSW. The lakes are ecologically, scientifically and recreationally important and are considered a permanently inundated water lacustrine system. The last few decades have resulted in observable drying trend visible throughout many of the individual lake systems. This provides an important question: is the recent water loss in the lakes of Thirlmere Lakes National Park (TLNP) a result if a degraded environment in the TLNP, or are the lakes just sensitive to natural hydroclimate variability. This research report will aim to address this question by assessing the past hydrological variability of Lake Baraba in the TLNP, in order to understand the palaeo-environment of Lake Baraba and put into perspective the recent trend in water loss in the TLNP.

Understanding the palaeo-environments of Lake Baraba was achieved by broadly investigating the sedimentary and stratigraphic characteristics of Lake Baraba, examining the grain size, age, organic content and the elemental composition of the sediments. The results of these analyses were synchronously compared to one-another and compared to the context of other palaeo-environmental publications from south-east Australian palaeo-climate and lake level research throughout the late-Quaternary, particularly the Holocene.

The results found that within the last ~35 ka, Lake Baraba has undergone significant hydrological and sedimentological change, however throughout the Holocene (~12-0 ka), there was a marked shift to stable lacustrine conditions in Lake Baraba, resulting in a constant peat formation record since ~ 9.3 ka years cal. BP. While there is some evidence of drying phases on the lake margin during the mid-Holocene, these events are only minimal and are not recorded in the centre of Lake Baraba. Additionally, the research found apart from the evidence of pre-Holocene peat development (~35 ka cal. years BP) and their close proximity, Lake Baraba and Couridjah highlight different hydrological and sedimentological conditions inferring that they currently operate as independent water bodies, however they may become connected via surface water flow during periods of high rainfall. Therefore, these finding suggest that the Thirlmere Lakes are significantly unique within the Sydney Basin, presenting itself as a possible important palaeo-environmental archive throughout the late-Quaternary, in particular the Holocene in the context of south-east Australia.

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A degraded environment or an environment sensitive to natural hydroclimate variability?

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Cover Photo – is a juxtaposition of a historical photograph of Lake Nerrigorang, taken by Annete Bray in 1954, and a recent photograph taken in 2012 (Schadler and Kingsford (2016)).
ABSTRACT
The Thirlmere Lakes are a series of five freshwater lakes trapped in an ancient uplifted river meander, located on the eastern margin of the Greater Blue Mountains World Heritage Area, near Thirlmere, NSW. The lakes are ecologically, scientifically and recreationally important and are considered a permanently inundated water lacustrine system. The last few decades have resulted in observable drying trend visible throughout many of the individual lake systems. This provides an important question: is the recent water loss in the lakes of Thirlmere Lakes National Park (TLNP) a result of a degraded environment in the TLNP, or are the lakes just sensitive to natural hydroclimate variability. This research report will aim to address this question by assessing the past hydrological variability of Lake Baraba in the TLNP, in order to understand the palaeo-environment of Lake Baraba and put into perspective the recent trend in water loss in the TLNP.

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The results found that within the last ~35 ka, Lake Baraba has undergone significant hydrological and sedimentological change, however throughout the Holocene (~12-0 ka), there was a marked shift to stable lacustrine conditions in Lake Baraba, resulting in a constant peat formation record since ~9.3 ka years cal. BP. While there is some evidence of drying phases on the lake margin during the mid-Holocene, these events are only minimal and are not recorded in the centre of Lake Baraba. Additionally, the research found apart from the evidence of pre-Holocene peat development (~35 ka cal. years BP) and their close proximity, Lake Baraba and Couridjah highlight different hydrological and sedimentological conditions inferring that they currently operate as independent water bodies, however they may become connected via surface water flow during periods of high rainfall. Therefore, these findings suggest that the Thirlmere Lakes are significantly unique within the Sydney Basin, presenting itself as a possible important palaeo-environmental archive throughout the late-Quaternary, in particular the Holocene in the context of south-east Australia.
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1. INTRODUCTION;

1.1 Context

Wetlands and lake systems are indispensable environments contributing to a broad array of habitats and ecosystems providing for and benefiting the wellbeing of both natural and anthropogenic landscapes globally. A wetland is broadly defined as a landscape where the existence of water acts as the primary regulator of the environmental process, involving a large variety of environments from: all lake and river systems, coastal lagoons, mangroves and swamps, peatlands and man-made reservoir systems. Wetlands function within the environment to store and purify freshwater, mitigate flooding impacts and recharge underground reservoirs, while retaining nutrients to provide highly productive ecosystems, which act as a refugia for >40% of the world’s flora and fauna (Hu et al., 2017).

Unfortunately, wetlands continue to be lost, degraded and altered globally, more than any other ecosystem (Millennium Ecosystem Report, 2005). Natural flooding regimes are altered by interception of flow, degrading wetlands and effecting their dependant and abundant biodiversity (Schadler and Kingston, 2016). Increasingly, anthropogenic impacts associated with longwall coal mining are negatively impacting upon freshwater ecosystems, contributing significantly to pollution through both direct and indirect means whilst heavily impacting surface and groundwater flow regimes supplying significant wetland systems (Marandi et al., 2013; Schadler and Kingston, 2016).

The Thirlmere Lakes National Park (TLNP), consisting of five Lakes; Gandangarra, Werri-Berri, Couridjah, Baraba and Nerrigorang, is part of the Greater Blue Mountains World Heritage Area and is listed as a RAMSAR Wetland of International Significance. Lake water levels in the system have fluctuated throughout time, however recently water level fluctuations have become more prominent and significant, possibly linked to groundwater extraction in relation with nearby mining activity, which has become a primary concern within the wider local community.

To date, water level variability in the TLNP has been restricted by a lack of environmental data, particularly knowledge of the lakes geomorphic and hydrological history. Recent studies however have extended the historical records of estimated lake water levels (Black et al., 2006; Vorst, 1974; Barber, 2018) and investigated correlations between possible drivers of the increased water level variability. Schadler and Kingston (2016) found that declining water levels in the TLNP coincide with the establishment of longwall coal mining and the establishment of groundwater extraction from bores. These sources however noted that these effects were difficult to distinguish without a good understanding of the connectivity between surface and groundwater systems in the TLNP, which currently does not exist. There is still much to be discovered about the Thirlmere Lakes and their...
geology, geomorphology, hydrogeology and hydrology. Without this information the drivers of decreasing water levels within the lakes remains a mystery, making it difficult to determine and manage the primary drivers of surface water loss (Schadler and Kingston, 2016).

This Honours Thesis is part of a greater collaborative project between the Office of Environment and Heritage (OEH), University of Wollongong (UOW) and the Australian Nuclear Science and Technology Organisation (ANSTO).

1.2 Significance

1.2.1 Thirlmere Lakes National Park
The significance of the TLNP, its lakes and the environments and ecosystems are considered to be of high importance due to their unique geologic origins and biota (Horsfall et al., 1988).

- The TLNP is home to an endemic species of freshwater sponge (*Radiospongilla spectroides*), thought to only live within the lakes of TLNP and other restricted areas of the Warragamba Special Area (NPWS, 1979; 1995).
- The lakes environments provide important habitats for two waterbird species, the Australasian Bittern (*Botaurus poiciloptilus*) and the Japanese Snipe (*Gallinago hardwickii*), both of which are listed as endangered under the *Environmental Protection and Diversity Act 1999* (EPBC Act, 1999; Schadler and Kingston, 2016) and protected by migratory bird agreements (Schadler and Kingston, 2016).
- The formation of the TLNP between 15 and 2 million years ago occurred through a series of rare geological processes (Horsfall et al., 1988), adding that it is highly unusual for such small lakes to remain in existence for such long periods of time without becoming completely infilled. They exist today because of a small catchment area (4.6km²), in conjunction with highly resistant parent material (Hawkesbury Sandstone) and densely vegetated lake margins (Noakes, 1998; Timms, 1992; Pells consulting, 2011). Therefore they can give a very succinct local climate and paleo-history.
- The TLNP is made up of a series of perched lakes (all lying at different elevations), with Baraba being the highest in elevation.
  - Despite slow infilling rates, large alluvial fans have formed between each of the lakes, creating a series of five separate lakes (Riley et al., 2012).

1.2.2 Lake Baraba
Lake Baraba is described as a ‘perched lake’, meaning it exists at a higher elevation than the other lakes in the TLNP. This is highly unusual as Lake Baraba also demonstrates the highest modern water
level in the TLNP during the time of the study. Therefore the formation of Lake Baraba is highly interesting due to the well preserved records following the transformation of a former river meander into the current lacustrine peat forming environment of Lake Baraba. Initial indications highlight they possibility of a clay liner protecting the ‘perched’ nature of Lake Baraba, preventing water loss into the adjacent lakes Couridjah and Nerrigorang.

1.3 Aims and Objectives
The aim of this student project is to assess the past hydrological variability from Lake Baraba, to discern the palaeo-climate of the Holocene and better understand if the recent drying trend in the TLNP is unprecedented. This will be achieved through close examination of sedimentological, geomorphic and paleo-environmental characteristics from within Lake Baraba, and the sill separating Lake Baraba from Lake Nerrigorang.

1.4 Thesis Outline and Context
Following the introduction, this thesis presents a broad review of the relevant literature surrounding lake processes, sedimentary processes, peat, and recent changes in environment in south-eastern Australia throughout the late-Quaternary. Chapter 3 outlines the study area, modern geomorphology, hydrology and sedimentology specific to the Thirlmere Lakes. In Chapter 4, the methods undertaken to complete the data collection and detailed analysis are outlined. Following this, Chapter 5 presents the results of the study, through core descriptions, radiocarbon dating, Elemental analysis and organic composition of Lake Barba. Chapter 6 then discusses the results presented in relation to previous research and the broader context of south-eastern Australia, whilst Chapter 7 provides a succinct conclusion for the study and highlights some topics of recommendation for future research in the TLNP.
2. LITERATURE REVIEW

2.1 Lake Sediments

Lakes sediments accumulate throughout time. The rate of which they accumulate is broadly governed by influences from the local climate, lake morphology and the geomorphology of the surrounding catchment (Timms, 1992). Sedimentary inflow to a lacustrine environment can occur through stream, river or overland flows. These flows can deliver coarse grained, minerogenic sediment into a lake, which then disperses and is deposited after sorting by physical parameters of the lake, such as sediment size, lake depth, density and wave action (Timms, 1992). Generally, coarser sediments are deposited close to the shoreline of a lacustrine environment, whilst fine silts and clays separate from the coarse grains. The coarse grains settle quickly near the lake edge and fine grains towards the lake centre (Nelson and Lister, 1995). During dry periods, when the lake level drops, coarse grain material may be deposited closer to the lake centre, due to a relative shift in the lake margin. Therefore, low lake levels can be interpreted by a coarsening clastic sediment texture, with finer grained sediments signifying times of increased water levels. Therefore, in general, the spatial variability of sediment textures of a lacustrine environment can be used as a proxy to infer previously lake margins, sediment transport dynamics and biological processes that operate within the lake system (Nelson and Lister, 1995).

Closed lake systems such as the TLNP, with few inflowing streams, can exhibit predominantly autochthonous organic sediments, which closely reflect the biological characteristics of the lake environment. These sediments reflect the lake characteristics well, because the organic matter comprises of a complex mixture of biochemical, which are derived solely from organisms previously living within that lacustrine environment (Meyers and Ishiwatari, 1993).

2.1.1 Lake Sediment Studies (Palaeolimnology)

Lakes are sinks for both water and sediment, thus their composition and the character of their sediments has been recognised to hold the potential to reveal both the long and short term history of a lake, and its surrounding environment (Gergis, 2000; Leeder, 1999). Lake sediments preserve and provide what is mostly a full record of the medium to long term effect of natural and anthropogenic impacts on our environmental systems (Clark, 1997). The merits of lacustrine sediments lie in the physical, chemical and biological diversity of the inflow materials- allowing the past climatic conditions of the lake catchment to be interpreted (Batterbee, 1991; Gergis, 2000). Impacts such as: climate change, acid precipitation, eutrophication and atmospheric deposition of pollutants (i.e. mercury and organic pollutants) to be interpreted from a sedimentary sequence (Douglas, 2013).
Lacustrine sediments can be classified into three general classifications based upon their origins: autochthonous sediments are native to their location (e.g. carbonate precipitation), allochthonous sediments which refer to sediments physically transported from their previous location (fluvially transported clastic material), and para-autochthonous sediments which show mixed characteristics (Rubensdotter and Rosqvist, 2008; Huang et al., 2017). The origin of sedimentary material deposited within lacustrine records therefore provides empirical evidence to the links and interactions between anthropogenic and natural processes in the environment (Dixon, 1998). The continuous nature of lacustrine sediments allows environmental process’ and trends to be observed over a multitude of timescales (Batterbee, 1991), whilst Clark (1997, pp. 3) explains that ‘multidisciplinary studies of sediment... can yield integrated reconstructions of past changes’. This provides insights into the interacting effects of cultural, climatic and ecological impacts upon the landscape.

2.2 Lakes as palaeo-environmental archives

The accumulation of sediments within lacustrine environments includes the deposition of many palaeo-environmental proxies, which can be meticulously preserved within the lake sedimentary record. The nature of the proxies found within a lacustrine records, depends upon the basin morphology, hydrological characteristics, catchment size and its position in the landscape. Environmental proxies commonly used include charcoal, pollen and diatom analysis, which are often combined with geochemical data such as stable isotope composition (e.g., $\delta^{13}$C and $\delta^{15}$N) and elemental analysis. When combined, these data sets can give a detailed insight to aid the interpretation of palaeo-environmental records, in combination with the inclusion of geo-chronical techniques (e.g. Accelerated Mass Spectrometry (AMS) radiocarbon dating and Optically Stimulated Luminescence (OSL) to create accurate age-depth models (Donders et al., 2007).

2.2.1 Australian palaeo-environmental reconstruction studies

Mainland Australia contains relatively few lakes, with most being shallow and ephemeral, which experience dry conditions during long and extensive droughts (Bridgman and Timms, 2012). Despite consisting of a limited number of lacustrine environments, Australian lakes highlight a wide range of origins. Timms (1992) categorised Australian lakes by their origin into: Tectonic lakes, volcanic lakes, coastal lakes, lakes formed by landslides, glacial lakes and lakes formed by wind action.

Although few in number, lacustrine systems in south-east Australia have been extensively studied to infer palaeo-climatic shifts throughout history. South-east Australian lakes which have been studied in detail include the Coorong coastal lakes in South Australia, Maar lakes in south-western Victoria (Jones et al., 2001) and Lake George, one of Australia’s largest lakes storing the longest quasi continuous sedimentary archive of any Australian lake basin (McPhail et al., 2016). Detailed palaeo-
environmental reconstructions in the TLNP have been frequently conducted and by many recent authors; Clark (1997) used paleo-ecological techniques to understand previous vegetation dynamics and the impacts Europeans have had on the TLNP. Gergis (2000) described the postglacial environment of the TLNP using a sediment based study from Lake Couridjah, whilst Black et al. (2006) who created a >43000 year vegetation and fire record using sedimentary cores from Lake Baraba.

Black et al. (2006) used a palaeoecological approach to address palaeo-environmental, vegetative and bushfire management strategies implemented within the TLNP. Pollen was used as the main indicator of vegetative histories of the TLNP, as ratios of pollen abundance derived from key taxa *Casuarina* and *Myrtacae (Eucalyptus)* can indicate differences in key periods of arboreal conditions (forest expansion) vs wetland/swampy conditions (lake expansion) (Black et al., 2006; Donders et al., 2007). Thornhill (2012) further explains how changes in diversity and abundance of *Myrtacea* pollen can reflect different environmental states and hydrological variations in an environment. Black et al. (2006) concluded that high *Myrtacea/Casuarinaceae* ratios indicate periods of swamp like conditions, whilst low *Myrtacea/Casuarinaceae* ratios indicate a lake like environment.

2.3 Peat;

2.3.1. Peat formation

Peat accumulates in environments where the rate of organic material deposition exceeds the rate of microbial decomposition in that environment (Hope et al., 2009; Cowley et al., 2016); usually in a proportion of around 10% solid weight to 90% water weight (Cameron et al., 1989). The simplest physical requirements for peat formation were succinctly explained where highly organic sediment receives protection from both desiccation and decomposition under saturated conditions and most importantly, a constant high humidity level. Therefore, as a consequence of the specific conditions required for peat formation, the control mechanisms that govern peat formation greatly differ from those factors which control the minerogenic and fine organic material deposition. It is the requirement of constant, stagnant saturation meaning that peat formation closely associated with peat bog and mires, opposed to lakes and lacustrine systems.

In lacustrine settings, peat is most likely to form within shallow basins that have high nutrient concentrations and a discontinuous drainage pattern which leaves the sediment constantly saturated. Some lake systems, such as those found in the TLNP, contain prime peat forming conditions, thus they must exist somewhere on the geomorphic spectrum in between a lacustrine and peatland environment. Examples of other lacustrine-mire systems, such as the Tasik Bera in Malaysia are characterised as a peat forming environment due to limited drainage processes and an
elevated water table, which have formed in response to tectonic activity within the basin, or by the accumulation of organic matter damming the outflow point of the system (Phillips and Bustin, 1998).

2.3.2 Peat forming environments

2.3.2.1 Peatlands

Peat forming environments or ‘peatlands’ cover only a limited 5% of the Earth’s surface and are often regarded as harsh environments due to waterlogging, low nutrient input and acidic conditions in an anaerobic setting. Regardless of their limited prevalence, peat deposits of varying thickness and extent are common in many types of geological and geographical settings throughout the world. There is great complexity, with literature regarding different types of peat forming environments and their classifications. They are highly variable in their nature throughout different global environments and systems and specific classification frameworks have been developed and applied on a case-by-case basis to peat forming regions. The term ‘peatland’ and ‘mire’ are relatively interchangeable, as environments named as bogs, fens, swamps, moors and marshes all fall within the category of ‘peatlands’, however peatlands are usually differentiated according to the composition and interactions between vegetation, hydrology and water chemistry of a system (Moore, 1995; National Wetlands Working Group, 1997; Hope et al., 2009, 2012).

Table 2.1, indicates the different interpretations of peat forming environments throughout the world, demonstrating the importance of the study area, including your research aims and context are highly important factors when deciding which definition of a peatland is appropriate (Inisheva, 2006). Figure 2.1 (Moore’s (1995) peatland classification table), highlights the relationship between precipitation/evaporation ratio and the influence of groundwater in the environment as the primary drivers of peatland classification, highlighting the general consensus that peatlands can be simply described as organic wetlands, which have the ability to accumulate peat.
2.3.2.2 Upland swamps

Where the slope permits, some peat forming environments can occur on plateaus, headwater reaches and high-altitude alpine peat forming environments. These types of peatlands have been described within the literature as ombrogenous (rain-fed) mires (Bragg and Tallis, 2001), alpine mires (McGolne et al., 1997) and headwater blanket peats (Holden and Burt, 2002), all of which have an important role in catchment scale processes, including water purification, regulation and filtration. In southeast Australia, upland swamps (Freidman and Fryirs, 2015; Young, 2017) and dells (Young, 1982) are found in many headwater regions, however they differ due to their vastly different hydrology and geomorphic processes in conventional peatlands described in 2.4.2.1.

According to Fryirs et al. (2014), upland swamps (also known as Temperate Highland Peat Swamps on Sandstone (THPSS)) are a type of Australian mire, which exists upon low-relief plateaus of the south-east Australian highlands. THPSS are particularly concentrated throughout the Woronora Plateau and Southern highlands of NSW, whilst also covering over 2000km² of area in the Blue Mountains, NSW (Young, 1982; Fryirs et al., 2014; Kohlhagen et al., 2013) (Figure 2.2).

Many THPSS started developing during the Holocene in response to an increased rainfall following the climate amelioration post LGM. They commonly occur in topographically low headwater reaches of low order streams (Jenkins and Frazier, 2010), normally associated with gentle slopes and minimal

Figure 2.1: Moore’s (1995) classification of peatland forming environments and production based primarily upon their hydrology.
stream-power encouraged sediment (derived from the underlying Hawkesbury Sandstone) deposition blocking headwater drainage lines. This process causes discontinuous drainage patterns, in turn providing the right for waterlogging to occur, simultaneously promoting the accumulation of organic matter (Fryirs et al., 2014; Cowley et al., 2016).

Indeed, whilst THPSS of south-eastern Australia are similar to fens of the Northern Hemisphere in terms of water and carbon storage abilities, their underlying sandstone lithology makes them distinct to the south-eastern regions of Australia (Cowley et al., 2016). THPSS highlight prominent mineralisation within peat layers, with the overall swamp substrate typically containing a mixture coarse-quartzose sand and peat which is usually connected to an overlain peaty mat at the sedimentary surface (Jenkins and Frazer, 2010).

![Figure 2.2: Temperate Highland Peat Swamps on Sandstone across NSW. The data freely available from NSW Office of Environment and Heritage website (Map source: Barber (2018)).](image)

### 2.3.3 Peat classification

Geologically speaking, peat is very young (a few hundred thousand years old); forming mostly during the Quaternary, with the type of plant, and the degree in which it has decayed is a major control of the organic composition of peat (Cameron et al., 1989).

Peat formations can be split into three different sub-types (Fibric, Hemic and Sapric) depending on their composition and stage of decomposition (Simonson, 2018; Cameron et al., 1989). Fibric peat is characteristic of intact and undecomposed plant material, which yields predominantly clear water. Hemic peat may contain some recognisable plant matter (roots, stems, leaves, wood, bark and...
seeds), and express water laden with mud when compressed, often reddish brown/brown in colour (Cameron et al., 1989). Sapric peat is heavily decomposed, with no intact plant remains, and remains a dark oozing mud upon squeezing a sample between one's fingers (Young, 2017; Cameron et al., 1989). The different peat types are also often characterised by their depth with a peatland formation. Fibric peat is often found in the upper, vegetated layers (acrotelm), whilst Hemic and Sapric peat facies exist deeper down, in the more anoxic sub terrain facies (catotelm), meaning that the portion of fibrous material within a peatland decreases with depth below the surface (Hope et al., 2009).

Table 2.1: Different classifications of peat forming environments from Australia, and around the world. Source: Barber (2018).

<table>
<thead>
<tr>
<th>Reference</th>
<th>Classification Description</th>
<th>Study area</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fryirs, 2016</td>
<td>Used the ‘River Styles Framework’ (see Brierly and Fryirs, 2005). A range of geomorphic indicators are used to assess ‘intact’ and ‘channelized’ swamps and their condition (good, moderate, and poor).</td>
<td>Blue Mountains National Park, NSW</td>
</tr>
<tr>
<td>Hope et al., 2009, 2012</td>
<td>Peatlands indicate terrestrial sediments in which organic matter exceeds 20% dry weight and with a depth generally greater than 30 cm.</td>
<td>Snowy Mountains, NSW and Australia Capital Territory (ACT)</td>
</tr>
<tr>
<td>Inisheva, 2006</td>
<td>Suggests peat types and peat soils should be classified based on floristics and botanical composition of the peat. Degree of decomposition and ash content are also important determinants.</td>
<td>Russia</td>
</tr>
<tr>
<td>The National Wetlands</td>
<td>Peatlands contain more than 40 cm of peat accumulation on which organic soils develop.</td>
<td>Canada</td>
</tr>
<tr>
<td>Working Group, 1997</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Moore, 1984, 1995 (Fig. 2.1)</td>
<td>Mires can be classified based on a range of different criteria including floristics, morphology and peat characteristics (see Moore, 1984). However, classification based on hydrological aspects is the most useful; groundwater or flow-fed (rheotrophic) and rain-fed (ombrotrophic) peatlands.</td>
<td>Europe</td>
</tr>
</tbody>
</table>
Hope et al. (2009) states that caution should be taken however, as heavily fibric peat facies can exist below humic facies. An example is emphasised by Vorst (1974), who describes a highly humified peat layer beneath layers of clay and sand within the Thirlmere Lakes sedimentary sequence. The inclusion of highly humic layers beneath non-organic sediments highlights the influence of strong periods of rapid peat growth, and irregular wetting/drying cycles within a wetland/mire can have within a peatland undergoing complex climatic variation (Kalnina et al., 2014).

2.3.4 Peat humification

Peat humification is the process by which remains of both flora and fauna (peat) are transformed into complex and heterogeneous mixtures of polydispersed materials known as humic acids (Zaccone, 2018). Understanding the evolution of organic material into humic acid in peatlands is of key importance to understanding soil carbon geochemistry and the ecosystem services which are provided by organic soils and wetlands, such as carbon sequestration and climate change mitigation (Zaccone, 2018).

Although vegetation growth and supply in a peatland may be extensive, peat formation is dependent on exterior factors such as moisture level and nutrient availability. Changes to the moisture or nutrient availabilities in a peatland can directly affect the net carbon accumulation of an environment, indirectly altering the factors governing the slow process of organic decomposition; oxygen availability, temperature, water saturation, litter availability and pH of the environment (Zaccone, 2018).

Atomic ratios are the most common humification proxy used in palaeo-environmental and geochemical studies, with the C/N ratio one of the most common proxies which can be used to describe humification patterns in contemporary peat profiles (Zaccone, 2018). This technique relies on the observation that during the decay of organic material, there is enrichment in Nitrogen in comparison to Carbon levels within the peat, thus resulting in a decrease in the C/N ratio of that peat.

2.3.5 Lake and Peatland Evolution

Because lakes continually accumulate sediments, they often slowly transform over time as sediment levels build up, becoming shallower with a smaller surface area (Timms, 1992). This process of infilling in lacustrine environments is referred to as ‘hydroseral succession’ or ‘terrestrialisation’, a process normally taking hundreds to thousands of years to develop (National Wetlands Working Group, 1997; Heathwaite, 1993).
Terrestrialisation occurs within lacustrine environments from a regime of significant infilling, transforming a lake into a terrestrial environment. The process of terrestrialisation is the main origin of mires and bogs throughout the Northern Hemisphere, which typically involves a shift from lacustrine sediments to increasingly organic peat material (Tallis, 1973; Balaga, 2007). Moss et al. (2016), found that terrestrialisation has also occurred in Australia, specifically Fen complex’s on Fraser Island, south east Queensland, which developed due to gradual infilling of deep lake systems, noted at many sites through the clear transition from a deep lake, towards shallower water after infilling, before the establishment of peat forming rush *Empodisma minus*.

Within small steep sided lake systems, or perched basins influenced heavily by climatic fluctuations, the process of territorialisation can occur in the presence of a ‘schwingmoor’ structure, where rapidly increasing water levels can completely flood a peat mire complex, which results in the uplift and floating of peat rafts and islands (Tallis, 1973). After floatation, the basal layers of the peat raft then settle to the lake floor, actively infilling below the peat islands and rafts.

**Floating peat islands;**
Vorst (1974) and Fairley (1978) reported the presence of large floating islands consisting of peat floating upon the lake surface, which in 1974 were between 0.5m diameter to over 5m across and a thickness of no more than 1m (Vorst, 1974). The appearance of the islands vary, with some bearing the remains of aquatic sedges and lilies, and melaleuca, whilst some islands are relatively bare, displaying only herbs and grasses. This variation of plant flora species on the floating peat islands suggests that they have been floating on the surface of the lakes for different periods of time.

The peat islands all tend to congregate at the northern and southern edges of the lakes, normally along the line of the greatest wind fetch on the lake-ward side of the reed margin, as they display ‘iceberg properties’, meaning that 95% of the islands mass is below the water surface, making them relatively easy to manoeuvre (Vorst, 1974). The presence of peat islands provides valuable information explaining the sedimentary processes occurring within the TLNP. It has been discovered that the islands are made from materials which are ‘almost identical’ to the sediment which is found on the lake floor (Vorst, 1974, p. 34).

2.4 Late Quaternary environmental change in southeast Australia;

2.4.1 Early Last Glacial Period (~35-22ka)
The early glacial period in south-eastern Australia coincides with a period of moist conditions with high fluvial activity, before evolving into a cool dry climate with the onset of the subsequent glacial period (Petherick et al., 2013; van Meerbeeck et al., 2009; Markgraf et al., 1992; Harrison and
From ~35ka onwards, high effective moisture was indicated by increased fluvial activity in the Murray Darling Basin (Page et al., 1996), and in Lake George, which showed periods with water depths up to 37m (Coventry, 1976). Approaching ~32ka (Headley Tarn Advance), glaciers started to occur in the Snowy Mountains (Petherick et al, 2013) and arboreal taxa started to be replaced by herbaceous species more acclimatised to the cooler, dryer south-eastern climate (Petherick et al., 2011).

Evidence of higher sediment loads were observed throughout many of the tributaries feeding large south-east Australian rivers; The Lachlan, Murrumbidgee, Murray and Goulbourn Rivers all developing large paleo channels in their tributaries, whilst palaeo-channels from the Murray Darling Basin (MDB) have been dated at 35-25ka (Petherick et al, 2013). Estimates in the MDB, measuring the meander wavelengths and reconstructed channel cross-sections indicate that bank-full discharges were 5-10 times larger than present between 35-25ka (Petherick et al., 2013), whilst similar records were found throughout the Riverine Plains during this time period indicate lakes fed by major river systems were deep or overflowing (Petherick et al., 2013).

High sediment loads were driven by newly denuded and destabilised catchment headwaters with the onset of glaciation (~32ka) (Petherick et al., 2013). This increased fluvio-lacustrine activity represents increased catchment run-off, a product of increasing snowmelt in the highland headwaters and reduced vegetation cover during the early Last Glacial Period (Petherick et al., 2013).

2.4.2 Last Glacial Maximum (~22-18ka)

The climate of the Last Glacial Maximum (LGM) was characterised by increased aridity, evapotranspiration and wind, experiencing less than half modern precipitation levels of south east Australia (Petherick et al., 2011; Gergis, 2000). During the LGM, the Australian landmass was roughly one-third larger due to eustatic sea levels 120m lower than present (Petherick et al., 2008). As a result, the lake levels of south-eastern Australia were low (Petherick et al., 2011), in a mostly treeless landscape above 600m AHD (Petherick et al., 2011; Kemp and Hope, 2014). Whilst the majority of palynological records indicate dominant open steppe grassland featuring species such as Poaceae, Asteraceae and Chenopodiaceae (Petherick et al., 2011), others indicate the opposite (Black et al., 2006). From a pollen record collected from Lake Baraba, Black et al. (2006) depicted Casuarinaceae dominance during the late glacial period, suggesting the TLNP may have acted as a refugium for arboreal species.
2.4.3 De-glacial Period (~18-12ka)

The general increase in temperature and moisture during the subsequent glacial-interglacial transition (~18ka), is indicated by the significant and rapid retreat of glaciers in the Snowy Mountains (Barrows et al., 2001) coupled with the re-emergence of arboreal species throughout temperate south-eastern Australia (Williams et al., 2009; Petherick et al., 2013). By 16ka, de-glaciation was virtually complete, apart from short-lived readvance at Mount Twynam, Snowy Mountains and Tasmania 16.8ka (Petherick et al., 2013).

In south-eastern NSW however, the de-glacial period remained relatively arid (Kemp and Hope, 2014) highlighted by lake levels in Lake George falling significantly lower ~14-10ka (Fitzsimmons and Barrows, 2010).

2.4.4 Holocene (~12-0ka)

Throughout south-eastern Australia, the onset of the Holocene (~12ka) brought increased temperatures and precipitation with a generally stable climate (Kemp and Hope, 2014) reaching hypsithermal (Holocene thermal maximum). Petit et al (1999) collected an ice core spanning 420 ka of Antarctic climatic history and found the current climate of the Holocene to be the longest warm period throughout the last four interglacial cycles. The increased effective moisture is evident throughout south-eastern NSW, e.g. Mountain Lagoon (Blue Mountains, NSW) persisted as a swamp until becoming a lake ~10ka and Lake George experienced its highest Holocene water levels ~10-8ka (Fitzsimmons and Barrows, 2010). Furthermore, mires at Barrington Tops started to develop peat from 8.6 ka (Dodson, 1987), and Lake Baraba 8.5ka (Black et al., 2006) highlighting swamp vegetation and permanently saturated conditions throughout parts of south-eastern Australia during the early Holocene.

Early Holocene palaeo-ecological records indicate increased forest cover throughout the semi-arid to temperate sub-humid regions of south-eastern Australia throughout the early-mid Holocene (Fitzsimmons and Barrows, 2010).

Towards the mid-Holocene (~6ka), south east Australia reverted back towards drier, more arid climate (neoglacial) characterised by greater climatic variability (Dodson, 1987; Harrison, 1993; Woodward et al., 2014; Tyler et al., 2015), thought to be due to the onset of modern climatic processes, such as increased Walker circulation (Harrison and Dodson, 1993), the primary driver of El Nino Southern Oscillation (ENSO) patterns (Power and Smith, 2007). Although open system throughout south-eastern Australia suggest a relatively drier climate between 8-6ka, Jones et al. (2001) and Sweller and Martin (2001) both found evidence suggesting wetter conditions across south-east Australia in closed systems (Fitzsimmons and Barrows, 2010). El Nino ENSO conditions are
dominant, which induce pronounced dry periods throughout south-eastern Australia (Woodward et al., 2014; Reeves et al., 2013).

During the late Holocene (~4ka till present), the climate of south-eastern Australia has experienced more variable climatic conditions. Palynological records show Lake George exhibiting hydrologic variability through intermittent pulses of lake transgression in the last 2.5ka (Fitzsimmons and Barrows, 2010), intermittent glacial fluctuation in the Snowy Mountains (Gergis, 2000; Gellantly et al., 1988), while in the Australian Capital Territory, mires containing peat show sand lenses around 3.5ka, after which peat is overlain indicating differing catchment driven by responses to the variation (Hope et al., 2009). Late-Holocene climate change, although weaker in amplitude than the dramatic climate shifts experienced during the last glacial cycle, have been shown to be increasing in its effects and frequency than commonly recognised throughout the Quaternary (Mayewski et al., 2004), and has subsequently influenced environmental change (Silva-Sanchez et al., 2014).
3. ENVIRONMENTAL SETTING

3.1 Study area

The Thirlmere Lakes National Park (TLNP), located 93 kilometres southwest of Sydney (Figure 3.1) is described as an upland fluviatile system consisting of five freshwater lakes; Gandangarra, Werri-Berri, Couridjah, Nerrigorang and Baraba, contained within an entrenched meander at an altitude of 305m above sea level (ASL) (Timms, 1992). Tectonic activity (explained in 2.2.1 Evolution of the Thirlmere lake sequence) beheaded a river that is believed to have previously flowed westwards, resulting in the isolated sinuous river channel (Timms, 1992; Black et al., 2006). The now closed, individual lakes act as efficient sediment traps that potentially document a history of environmental conditions (Vorst, 1974; Horsfall et al., 1988), which Gergis (2000) argued could be 15 million years long.

The 630 Ha Thirlmere lakes National Park belongs within the wider Nattai Reserves System, one of the largest conservation areas Australia, containing an undisturbed record of the environmental history for the late Quaternary period in the TLNP (Gergis, 2000).

Figure 3.1: The regional position of the TLNP within the Sydney Basin. The insert map shows a satellite image of the five lakes within the TLNP (source: Barber 2018).
3.1.2 Climate
The Thirlmere Lakes National Park has a warm temperate climate with temperatures ranging between 25 to 33°C in summer and 5 to 15°C in winter (Black et al., 2006). Horsfall et al. (1988) reported a mean lake temperature of 20°C, with a seasonal range of 28°C to 11°C, adding that average annual long term rainfall is 804mm (BOM, 2017 with bimodal wet periods between January-March and June (OEH, 2016). The wind regime consists of predominantly north-easterly winds in summer and south-westerly in winter (BOM, 2017).

3.1.3 Geology
The geology of the TLNP is representative of the southern Sydney Basin sequence, comprising of the Triassic Narrabeen Group characterised by a number of different formations and sub-groups; including the Bald Hill Claystone. The regionally important Bald Hill Claystone is part of the Clifton subgroup (a low-permeability, highly weathered Permian volcanic stratum) (Branagan et al., 1973; Schadler and Kingston. 2016), existing >100m below the lakes surface (OEH, 2016). Overlaying the Bald Hill Claystone, the Hawkesbury Sandstone is described as medium to coarse grain fluvially derived quartz sandstone, containing minor shale and laminite lenses, which tend to be sub-angular and sub-rounded in shape (Gergis, 2000). The Hawkesbury Sandstone is the dominant lithology of the TLNP (Figure 3.3) and forms the outcrops of erosion resistant ridgelines and steep valleys of the Thirlmere Lakes system (Fanning, 1982). Overlaying the Hawkesbury Sandstone, the Wianamatta group is the youngest member of the group comprises of mainly shales, laminate and sandstone units. Classed under the Wianamatta group, these sub-units including the Ashfield Shale, Minchinbury Sandstone and Bringelly Shale are also seen in the TLNP. Therefore due to the resistant geology and small nature of the TLNP catchment area, aging in the lacustrine system of the TLNP has occurred very slowly hence recording a long record of previous climate history within its sedimentary facies.
3.1.4 Evolution of the Thirlmere lake sequence;

The Thirlmere lakes lie unusually ‘perched’ above the Burragorang Valley to the west and the Cumberland Basin to the east (NPWS, 1995), in their own intriguing and seemingly misplaced sinuous river channel. The origin of the Distinct ‘U’ shape form of the Thirlmere Lakes sequence is a product of a previously meandering river channel that incised into the Hawkesbury Sandstone complex (Timms, 1992; Gergis, 2000). It is widely accepted that the river was uplifted and subsequently cut-off from its previous south-west drainage pattern as a result of structural forces associated with the formation of the Lapstone Monocline and Southern Lapstone Structural Complex (Fanning, 1982; Black et al., 2006; Timms, 1992; Noakes, 1998; NPWS, 1995), during the mid-Cenozoic ~15 MYA (Fergusson et al., 2011). Strong east-west contraction in the TLNP and neighbouring regions caused high-angle, west-over-east reverse faulting patterns, which are currently visible as a series of fault zones which run across the Sydney basin; Victoria Park, Bargo and Picton Fault Zones, with the Thirlmere Lakes possibly occupying the ‘hanging wall’ of one of these faults (Fergusson et al., 2011). The accumulation of sediment in the Thirlmere Lakes began following the initiation of uplift in the region (Vorst, 1974) with sedimentary processes turning the previously high energy, fluvial system to the current lacustrine environment present in the TLNP. The longevity of the Thirlmere Lakes can be attributed to the highly erosion-resistant properties (high quartz content) of the Hawkesbury Sandstone, and a low rate of erosion (Geoscience Australia, 2017; Horsfall et al., 1988), as ordinarily, lake basins of this age would have underwent terrestrialisation by slowly-infilling over time (Timms, 1992).
3.2 Modern Geomorphology and Sedimentary History

3.2.1 Geomorphology

The small catchment area of the Thirlmere Lakes (4.6 km$^2$) in conjunction with the erosion resistant Hawkesbury Sandstone and densely vegetated margins of the TLNP leads to a slow rate of infill, resulting in a highly stable lake geomorphology with steep banks and flat bottom (Vorst, 1974; Noakes, 1998; Timms, 1992; Pells Consulting, 2011). During periods of fire however, transport material is likely increased due to disruption to the dense, natural vegetation found in the catchment, potentially increasing the suspended sediment loads within the lakes for up to several years (Smith et al., 2012).

Despite the slow rate of infill and sedimentation within the lake system, a series of alluvial fans (Figure 3.4) have formed on the lake margins. The sill formations are a product of the ephemeral tributaries in the TLNP transporting sedimentary material from the surrounding catchment. The formation of these five alluvial fans has produced a series of five separate lakes within the Thirlmere Lakes system (Riley et al., 2012). The sediments of the fans are Bimodal, consisting of fine-to-medium grained sand and silty/clay (Vorst, 1974; Barber, 2018), directly reflects the lithology of the Hawkesbury Sandstone parent material with a medium-to-poor drainage (Rose and Martin, 2007).

Within the alluvial fans, the sediments are poorly sorted with no distinct sorting down-fan (Vorst, 1974). The occurrence of the raised sill indicating that the fans are being deposited quicker than the system’s ability to erode these sediments, which is typical of a generally low-energy state found in lacustrine systems (Timms, 1992).

![Figure 3.4: The location of alluvial fans infringing on the valley floor of the Thirlmere Lakes (numbered 1-5 from upstream). The black line represents the confines of the Thirlmere Lakes catchment area (4.6 km$^2$). The purple dotted lines indicate alluvial fans. The approximate areas of the alluvial fan catchments are as follows; 1 = 0.15 km$^2$, 2 = 0.24 km$^2$, 3 = 0.54 km$^2$, 4 = 0.08 km$^2$ and 5 = 0.21 km$^2$. The approximate areas of the alluvial fan catchments are as follows; 1 = 0.15 km$^2$, 2 = 0.24 km$^2$, 3 = 0.54 km$^2$, 4 = 0.08 km$^2$ and 5 = 0.21 km$^2$. (Source: Barber, 2018).](attachment:image.png)
3.2.2 Sedimentary History

Fanning (1982) stated that alternating layers of organic (Black/grey) and inorganic bands (yellow/orange/red) have been identified in the Thirlmere Lakes, suggesting that they may represent both open and closed lacustrine environments. Alternatively, Noakes (1998) hypothesised that these alternating bands may just represent changing water depths in the Thirlmere Lakes (Atkinson, 2000; Black et al., 2006; Schadler and Kingston, 2016). Vorst (1974) on the other hand suggested that the change from predominantly inorganic to organic sedimentation was due to the colonisation of the shallow water environments by aquatic vegetation.

Black et al. (2006) summarised the sedimentary history of Lake Baraba providing an age/depth relationship for the lake centre (Figure 3.5). Lake Baraba was initially a lake slowly accumulating clays (~0.04mm/y), during Marine Isotope Stage (MIS) 3 (~59-24ka), alternating between Yellow/red/orange bands indicating oxidising conditions (605-635, 510-535 and 410-262cm, possibly reflecting dry periods of the lake) whilst uniform dark organic and grey clays indicate anaerobic conditions indicating permanently wet lake conditions. Thus it can be discerned that both drier and wetter conditions alternated dominance in MIS 3, prior to the LGM, however just how ‘dry’ or ‘permanent’ the lake was prior to the LGM is difficult to assess (Black et al., 2006). In the Early Holocene (~8.5 ka), peat started to accumulate and the sedimentation rate markedly increased (~0.67mm/y) (Black et al., 2006), indicating that infilling had created water levels shallow enough to facilitate aquatic vegetation and peat accumulation. The accumulation of peat has continued until the present, mainly through coprogenic littoral peat development, which is interbedded with some alluvial sedimentary deposits (Timms, 1992; Gergis, 2000). A hiatus in the peat record was highlighted by Black et al. (2006), between ~6-5.2ka, perhaps in response to a dry period while Pickett et al (2004) reported increased moisture stress in mid-Holocene vegetation (between 6-5ka).
Figure 3.5: Age depth model for Lake Baraba created by Black et al. (2006).
3.3 Lake Hydrology, Limnology and Groundwater

3.3.1 Hydrology
The TLNP is predominantly rainfall fed (NPWS, 1997; Gergis, 2000; Horsfall et al., 1988; Schadler and Kingston, 2016; Russel et al., 2010; Riley et al., 2012) with lake water predominantly serving to recharge the aquifer system connected below (Pells, 2011; Riley et al., 2012; Schadler and Kingston, 2016). Surface water occasionally overflows from the lakes towards the west through Blue Gum Creek during periods of high rainfall, before flowing into the Little River in the Nattai National Park (NPWS, 1995; Horsfall et al, 1988). Bowler (1981) developed a hydrologic classification table of Australian Lake Basins using both lake morphology (lake and catchment area) and climate (evaporation, precipitation, and run-off co-efficient) datum to categorise Australian lakes. According to Bowler’s table, the Thirlmere Lakes are found on the boundary between permanent and ephemeral lake conditions, but on the ‘wet’ (left) side of the hydrologic threshold (Figure 3.6). This classification highlights the Thirlmere Lakes are positioned both morphologically and climatically to experience perennially wet conditions, yet they may also experience intermittent dry periods. The morphometric variables Bowler (1981) used, were derived from the lake and lake catchment area estimate (0.46 km$^2$ and 4.5 km$^2$), derived from the Thirlmere Lake Enquiry (OEH, 2012), whilst the climate values were derived from BOM (2017) datum, collected from the Picton weather station.

![Figure 3.6: Lake Classification showing the characteristics of Thirlmere Lakes by comparison to other major Australian Lakes (Bowler 1981). The Y-axis indicates catchment morphology, where Ac = Catchment Area, Al = Lake Area. The X-axis indicates lake Climate Function, where Fc = \((0.8E - P/Pf) + 1\), where E = evaporation, P = precipitation and Pf = run-off coefficient x P. Note two data points are plotted for each lake using two different runoff coefficients, (0.1 and 0.3). This is because generally Australian lakes basins have a run-off coefficient that rarely surpasses 10%, but during periods of heavy rain the run-off coefficient can achieve 30%. Steady state conditions are defined by the ‘hydrologic threshold’. (Source: Barber, 2018).](image-url)
3.3.2 Limnology

The Thirlmere Lakes are classified as hypereutrophic systems, meaning they experience high productivity with dense algal and macrophyte populations with limited light penetration of the water column. This hypereutrophic classification reflects only the recent conditions experienced in the Thirlmere Lakes, as Horsfall et al. (1988) highlighted the lakes were previously meso-oligotrophic in nature with slightly acidic waters that are low in calcium carbonates, producing visibly clearer water during period of higher rainfall, (Schadler and Kingston, 2016).

All five lakes share similar physical and chemical conditions; however each lake does show some slight variation. All the lakes are slightly acidic, with a dark brown water colour resulting from humic acids leaching from within the peat deposits below, with Couridjah presenting the darkest colour and highest turbidity count (Horsfall et al., 1988).

3.3.3 Groundwater

Residing directly below the TLNP are two main underground aquifers systems: one shallow and one deep which are separated by a layer of low permeability (Figure 3.7) (Branagan et al., 1973; Herbert and Shelby, 1980; Pells and Pells, 2012; Schadler and Kingston, 2016). The sub-surface alluvial aquifer contained within the relatively unconsolidated alluvial valley fill, is recharged by rainfall seeping down into the shallow aquifers underlying the Thirlmere lakes system within the underlying porous Hawkesbury Sandstone (Riley et al., 2012). A series of small perched aquifers intermittently occur within the sandstone ridges surrounding the sinuous river channel, slowly discharging as spring flow into the lakes (Russel et al., 2010), although very little is known about the groundwater processes in the TLNP. The deep aquifer in the TLNP occurs within the Narrabeen Group, and has no interaction with the hydrology of the lakes, as it is divided from the intermittent aquifer by the Bald Hill Claystone, which is a regionally significant aquitard, and does not allow any interaction between the Hawkesbury Sandstone and Narrabeen Group (Russel et al., 2010).

Schadler and Kingston (2016) hypothesised a low permeability Bald Hill Claystone layer prevented interaction between the deep and shallow aquifer below which has since been interrupted by anthropogenic events causing water loss from the TLNP. Investigations by the NSW Office of Water found declining water levels in the TLNP were primarily due to climatic variability due to insufficient rainfall (Russel et al., 2010), while a NSW government commission of inquiry concluded there was no strong evidence for anthropogenic influences accelerating the decline of water level in the TLNP. Pells (2010) disagreed with this conclusion, who estimated the lake water levels are 1.5-2.5m lower than the expected water level if natural climate variability was the primary driver of the declining lake level in the TLNP (Riley et al., 2012; Schadler and Kingston, 2016).
Rainfall throughout eastern Australia is known to be highly variable (Riley et al., 2012). Schadler and Kingsford (2016) highlight that water levels in the Thirlmere Lakes are strongly correlated with rainfall over a centennial timescale (1980-2015). In summary, a rainfall deficit (a below average rainfall) of 1600mm over a 50-10 year span would cause extensive drying of the Thirlmere Lakes (Riley et al., 2012). Low water levels have been recorded at the Thirlmere lakes throughout the 1937-1947 “WWII drought” (Schadler and Kingsford, 2016), while above average rainfall during 1974 caused the lakes to flood and overflow each corresponding sill (Vorst, 1974).

3.5 Flora and Fauna
There are over 400 species of flora found in the Thirlmere lakes National Park, representing over 250 genera (Benson and Howell, 1994; NPWS, 1995). Of these, only 13 plant species are considered non-endemic (NPWS, 1995; Gergis, 2000).

3.5.1 Terrestrial flora
The flora of the TLNP is typical of that throughout the Hawkesbury Sandstone complex (Benson and Howell, 1994). The catchment is dominated by dense, dry sclerophyll woodland and forest with a mixed composition of *Eucalyptus/Corymbia/Casuarina*, particularly Red Bloodwood (*Corymbia gummifera*) and Sydney Peppermint (*Eucalyptus piperita*) (Black et al., 2006) (Figure 5.8). Introduced species such as Pinus radiate and ornamental trees, shrubs and grass are now common species in the TLNP, originally planted on private land (Wollondilly Shire Council, 1965; Noakes, 1998).
3.5.2 Aquatic flora

The vegetation is dominated by shallow sedges which enjoy the infilled nature of the TLNP (Black et al., 2006; Vorst, 1974). Located right at the southern extent of its geographical range (Fairley, 1978; NPWS, 1995; Gergis, 2000), tall reed (*Lepironia articulata*) is the dominant aquatic species growing in dense thickets, at a water depth 2-4m deep. Closer to the shore pithy sword-sedge (*Lepidosperma longitudinale*), scale-rush (*Lepyrodia muelleri*) and zig-zag bog-rush (*Schoenus brevifolius*) are found (Fairley, 1978; Gergis, 2000; NPWS, 1995) onto which *Melaleuca Linariifolia* is encroaching (Black et al., 2006).

*Figure 3.8*: Thirlmere Lakes National Park, highlighting the six different vegetation communities and cleared land (source: NSW Department of Environment and Heritage (2003).

3.5.3 Fauna
The lake systems also provide important habitat for two waterbird species, the Australasian Bittern (*Botaurus poiciloptilus*) and Japanese Snipe (*Gallinago hardwickii*), which are listed as endangered under the *Environmental Protection and Diversity Act 1999* (EPBC Act, 1999; Schadler and Kingston, 2016) and protected by migratory bird agreements (Schadler and Kingston, 2016).

3.6 History;

3.6.1 Fire History
The TLNP experiences a major bushfire every three years (Noakes, 1988), supported by the visual evidence of charcoal in the subsurface layers (Black et al., 2016). The fire regime of the TLNP has been highlighted as one of the possible agents responsible for the initiation of sediment transport on the surrounding hillslope and alluvial fans, resulting in the large sills (Gergis, 2000; Vorst, 1974). Identification of substantial charcoal units in the subsurface of the hillslopes fans further suggest that fire was a main agent in the unusual fan and sill formation present in the TLNP (Vorst, 1974; Dixon, 1998; Gergis, 2000).

3.6.2 Human development, agriculture and mining history

3.6.2.1 Development and agriculture
The first development within the lakes’ catchment area occurred during the 1830’s. Early land use in the TLNP included small scale dairying, wheat farming and occasional timber harvesting (Fairley, 1978; Gergis, 2000). Mid-last century anthropogenic modifications to the lake system to benefit “power boating” included the removal of aquatic vegetation form lake margins, the construction of channel cuts between the lakes and the application of arsenic based herbicides (Lakes Werri Berri and Couridjah) (NPWS, 1979; 1995; Vorst, 1974). Power boating was restricted to only Lake Werri Berri after the lakes became a National Park in the 1970’s (NPWS, 1979; 1995; Vorst, 1974), before the TLNP was listed as a wetland of national importance in 2001 (Environment Australia, 2001) due to the large number of endemic and endangered species identified.

3.6.2.2 Mining
Many mining operations have been undertaken in the Thirlmere region; however the most significant to the lakes themselves is the start of longwall coal mining in 1982. Following 1982, groundwater levels in the shallow aquifer associated with the lake system fell by 40m (Schadler and Kingston, 2016), although there is no publicly available data on mining associated groundwater extraction. Schadler and Kingston (2016) identified drops in the water levels of three lakes in 1984, 1997 and 1998, identifying that these declines in water level were not climate related. Schadler and
Kingston (2016) suggest two hypotheses for this water drop; a) either considerably large amounts of water are being removed from the underground aquifer, or b) the presence of longwall coal mining in the area has disrupted underground regime of the TLNP.

Figure 3.10: a) the location of the TLNP in the Hawkesbury Nepean Catchment in NSW, b) other sites within the Greater Blue Mountains World Heritage Area (green outline) (T-Thirlmere, A-Avon Dam, N-Nepean Dam, W-Warragamba Dam, P-Picton), and c), where Blue Gum Creek (BG) takes water away from the five Thirlmere lakes within the boundary of the TLNP (Green area) in relation to the Tahmoor Colliery Lease (grey Area) and 20 of the nearby longwall mines (Dark Grey), with the date identifying the date each was established (source: Schadler & Kingston, 2016).
4. METHODS

4.1 Field Methods

4.1.1 Site Selection and Sub-surface data collection
The field work conducted for this project was focussed on the formation of Lake Baraba and its interactions with the sill separating it from Lake Nerrigorang to the north and Lake Couridjah to the south. All cores were extracted from the ground in aluminium pipes, which had a diameter of 74mm.

Figure 4.1: Exact location of all the cores along the LB-LN transect which were obtained during this study. Orange locations are Primary cores (BAR02, BAR03 & LBLNHA1, whilst Secondary cores (BAR04, BAR05 & LBLNHA2 are represented by blue location markers.
4.1.2 Lake and lake margin coring

One lake-centre core and three lake-margin cores were extracted from Lake Baraba using either a percussion corer or non-mechanical push coring method. BAR02 was taken from the centre of Lake Baraba (2.7m water depth) producing a 3.45m core. The core was extracted using a non-mechanical push core method, which was deployed from the edge of a floating peat island, with the assistance of a non-motorized boat.

BAR03 was taken from the contemporary eastern lake margin (60 cm water depth), with a non-mechanical push core method of retrieval, reaching a depth of 3.15 m. Cores BAR04 and BAR05 were taken from the northern lake margin using a percussion corer, BAR05 from right on the water’s edge and BAR04 was 3m north from the lake margin. Due to the extremely thick clay in the northern margins of Lake Baraba, the BAR04 and BAR05 were unable to be driven below 1m in depth.

4.1.3 Sill Sub-Surface Characteristics

Sub-surface investigation of the sill between Lake Baraba and Lake Nerrigorang was undertaken using a hand auger for the first 4m, before using a percussion corer for the last 2m (LBLNHA1), whilst LBLNHA2 (2.1m) and LBLNHA3 were removed using a hand auger. The sediment from the auger holes were documented in the field (colour, texture, sorting and major grainsize differences) with samples collected from every different stratigraphic unit.

4.1.4 Core Compaction

Core compaction occurred within cores BAR02, BAR03, ranging from 60-85 cm, which represents 17.4-26.9 % of core length, displayed below in table 4.1.

Table 4.1: Core compaction table

<table>
<thead>
<tr>
<th>Core code</th>
<th>Easting</th>
<th>Northing</th>
<th>Surface elevation (in AHD)</th>
<th>Method</th>
<th>Core length (m)</th>
<th>Core compaction (m), percent of total core length.</th>
</tr>
</thead>
<tbody>
<tr>
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<td>0273250</td>
<td>6209347</td>
<td></td>
<td>Push Core</td>
<td>3.45</td>
<td>0.6, 17.4%</td>
</tr>
<tr>
<td>BAR03</td>
<td>0273205</td>
<td>6209346</td>
<td></td>
<td>Push Core</td>
<td>3.15</td>
<td>0.85, 26.9%</td>
</tr>
<tr>
<td>BAR04</td>
<td>0273202</td>
<td>6209683</td>
<td></td>
<td>Percussion Core</td>
<td>0.77m</td>
<td></td>
</tr>
<tr>
<td>BAR05</td>
<td>0273204</td>
<td>6209749</td>
<td></td>
<td>Percussion Core</td>
<td>1.07m</td>
<td></td>
</tr>
<tr>
<td>LBLNHA1</td>
<td>0273202</td>
<td>6209751</td>
<td></td>
<td>Hand Auger/Percussion Corer</td>
<td>6.0</td>
<td>N/A</td>
</tr>
<tr>
<td>LBLNHA2</td>
<td>0273204</td>
<td>6209751</td>
<td></td>
<td>Hand Auger</td>
<td>2.12m</td>
<td>N/A</td>
</tr>
</tbody>
</table>
4.2 Laboratory Analysis

Surface and sub-surface samples were analysed both at University of Wollongong, Waikato University and ANSTO.

4.2.1 Subsampling
BAR02 and BAR03 were sampled at 10cm intervals, with approximately 1.5cm³ of sediment taken at each interval. The sampled sediments were bagged, labelled and then refrigerated (at approximately 4°C) until further analysis.

4.2.2 Grain Size Pre-treatment
Minerogenic material from very fine, cohesive and organic rich lacustrine sediments is traditionally hard to measure, because all organic material must first be removed (Vaasma, 2008). In cores BAR02 and BAR03, organic material makes up a large percentage of the sediment, therefore they had to be digested with 10% hydrogen peroxide solution (H₂O₂), to remove organic material. The samples were left to digest for approximately 9 days on a hotplate at 90°C, until there were no longer any visible signs of reaction. After digestion, the samples were placed in an ultrasonic bath, on high intensity for 15 minutes, to disperse any naturally occurring aggregates which may remain in the sample (Rendigs and Commeau, 1987).

4.2.3 Grain Size Analysis
The Malvern Mastersizer 2000 located at the University of Wollongong (UOW) School of Earth and Environmental Science (SEES) was used to determine the grain size distribution of the minerogenic material in the samples. The samples were added into a mixing beaker, flushed with tap water to remove residual H₂O₂ and then diluted to achieve adequate beam obscuration (10-20%). The instrument was flushed well after each measurement, to remove residual sediments before again measuring the next sample against the background solution.

4.2.4 Elemental Composition
The elemental composition of BAR02 and BAR03 was determined by an ITRAX micro X-ray Flourescense (XRF) core scanner housed in the Environmental Radioactivity Measurement Centre at ANSTO, Lucas Heights. A high-resolution optical image and radiographic image at a resolution of 500μm was obtained, performed using a molybdenum tube set at 30kV and 55mA and dwell time of 20s. A step size of 1000μm (1mm) was selected, capturing elemental variations occurring in the laminations visible in the radiographic image. The core scanner gives elemental composition of the core, but also includes ratios of incoherent and coherent scattering measurements (Inc/Coh), which can be used as a proxy for organic matter within lacustrine sediments.
Factor analysis by principle components, in correlation mode was applied to the physio-chemical properties of the ITRAX data, to synthesise the chemical compositions of the sediment and investigate the factors controlling the concentrations of the elements within the sedimentary cores. The subsequent ITRAX data was normalised to the Inc/Coh carbon ratio as to remove the influence of water content upon readings from the ITRAX results.

4.2.5 Organics analysis of %C, %N and C: N
The %C, %N and C: N ratio were analysed from BAR03 (36 samples) at 10 cm intervals. Once sampled, the sediments were transferred into labelled tin vessels before being dried in the oven at 50°C for 24 hours. The dry sediments were ground to homogenise the material. Once homogenised, the samples were weighed and pressed into tin capsules before being combusted in an Elemental Analyser (EA), with constant exhaust flow to a Nu Horizon isotope ratio mass spectrometer (IRMS).

4.2.6 Facies interpretation and cluster analysis
A cluster analysis was undertaken using PAST to determine the facies interpretation using the grainsize, standard deviation, skewness and kurtosis to determine the relationships between the sediments and facies found in Lake Baraba.

4.2.7 Accelerated Mass Spectrometry (AMS) Carbon-14 Dating
Three handpicked charcoal and plant macrofossils samples (2 from each core BAR02 and BAR03) were taken at selected depths and one bulk sediment sample (containing charcoal fragment) (BAR03) were dated using AMS techniques to obtain an approximate age chronology at selected depths with BAR02 and BAR03. Analysis was undertaken by the University of Waikato, Faculty of Science and Engineering Radiocarbon Dating Laboratory, NZ. All samples were pre-treated with an acid-alkali-acid treatment; samples first had to be cleaned and washed with hot HCL, before a second rinse with multiple hot NaOH solutions. The NaOH insoluble fraction from the sample was treated with hot HCL, filtered, rinsed and then dried. Once prepared, all samples were dated using Accelerated Mass Spectrometry (AMS). For a detailed description of the AMS dating process, refer to Beta Analytica Inc. (2018).
5. RESULTS

5.1 Core descriptions and sedimentology
Firstly, the core descriptions, grainsize data and Elemental analysis is detailed for the three primary cores, BAR02 (representing the centre of Lake Baraba), BAR03 (representing the margins of Lake Baraba) and LBLNHA1 (measuring the grainsize and lithology of the sill separating lakes Baraba and Nerrigorang). Then, the secondary cores, BAR04 (Lake margin/sill), BAR05 (northern margin of Lake Baraba) and LBLNHA2 (northern sill) is described in lesser detail (Figure 4.1). The interpretation and analysis of grainsize data was assisted by using Davies et al. (2015). Additional grainsize information can be found in the appendices.
5.1.1 Primary Cores

5.1.1.1 BAR02 – Lake Baraba centre core
BAR02 is a 3.45m continuous core which was taken from the centre of Lake Baraba (Figure 4.1) in water depth of 2.7 metres resulting in a core compaction rate of 17.39% (Table 4.1). The grainsize for core BAR02 is presented below, alongside sediment type composition and structure descriptions.

BAR02 (Figure 5.1) (modern lake centre) consists of very fine-to-medium grain silt (mean grainsize = 4.2-28.1 µm, silt composition =46-70%) characterised by a larger grainsize (20-30 µm) in the upper 0.3m of BAR02. Below 0.3m the grainsize drops <8 µm until two significant grainsize peaks at 0.3m (28.1 µm) and 2.3m (10.7 µm) depth (Figure 5.1). Silt and clay are the dominant sedimentary components of BAR02, ranging between (46-70% silt and 12-49% clay respectively).

Figure 5.1: (left to right) Optical image (taken from ITRAX), Sediment composition (blue = sand, red = silt, grey = clay), and grainsize (mean µm) for BAR02.

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Figure 5.2: Grainsize graphs for BAR02 from the Malvern Mastersizer 2000. (From top to bottom) A = 0.2m, B = 0.3m, C = 0.7m, D = 2.4m & E = 3.15m depth.
The upper 2.7m of BAR02 comprises of mostly homogenous organic material (peat), consisting of the semi-decomposing remains of decomposing plants and macrophytes, becoming increasingly humified below 1.5m. In the upper 0.4m (Figure 5.2a-b and Figure 5.3), the grain size is described as being a medium-coarse grain silt (mean grain size = 21.7 µm), which is very poorly sorted.

Below 0.4m depth, the grain size decreases to > 11 µm, becoming a poorly sorted-very fine silt (mean grain size = 6.3 µm) (Figure 5.2C and Figure 5.3). The grain size shows little variability between 0.4-2.4m with silt being the dominant sedimentary component (mean = 56%). At 2.4m depth (Figure 5.2D), there is a small-poorly sorted grain size peak (7.5 µm) coupled by a subsequent increased sand composition (sand = 14%) until 2.7m depth (Figure 5.1).

Below 2.8m, the peat becomes a dull grey, fine silty/clay unit, while the grain size remains relatively small (4.6-5.7 µm) until 3.1m (Figure 5.1). Following the interchanging peat/silt horizon at 3.2m, there is a second increase in grain size (11 µm) representing fine silt (Figure 5.5). At 3.3m (base of core), there is a relatively pallid clay plug (silt = 60.4% and clay =35.7%) with a grain size of 7 µm (very fine silt/clay).
Figure 5.5: Core photograph of sedimentary structure at 3.1-3.35m.
5.1.1.2 BAR03 – Lake Baraba western-margin core

BAR03 is a 3.15m continuous core which was taken from the modern western margin of Lake Barraba *(Figure 4.1)*, in a water depth of 0.6m with a 26.9% core compaction (table 4.1). The grainsize results of BAR03 are described below, alongside sediment composition and structure descriptions.

BAR03 *(Figure 5.3)* (modern lake margin) consists of a very-fine silt to fine sand (0.27-339 µm), with two major grainsize peaks at 0.6m (339 µm, *Figure 5.3a*) and 3.15 (742 µm, *Figure 5.3D*). Overall BAR03 demonstrates a coarsening grainsize down-core, which is reflected by the increasing sand composition down core, resulting due to the depletion of silts and clays below 1.3m within BAR03 *(Figure 5.2)*. 

*Figure 5.6*: (left to right) Optical image, Sediment composition (blue = sand, red = silt, grey = clay), Grainsize (mean µm) for BAR03.
Figure 5.7: Grainsize graphs for BAR03 from the Malvern Mastersizer 2000. (From top to bottom) A = 0.4m (peat), B = 0.6m (sandy lens), C = 2m (minerogenic sands), D = 2.8m (charcoal) & E = 3m depth (fine-grain sand).
The upper 1m of BAR03 consists of mostly homogenous organic material (peat) containing the remains of fibrous, semi-decomposed macrophyte remains becoming increasingly humified with depth (*Figure 5.8*). The upper 20cm of BAR03 is highly organic, resulting in a grainsize measurement for this section unattainable with our current methods (i.e. the weight of sample recovered after dehydration and digestion was too small to be analysed in the Mastersizer). There is a small peak in grainsize at 0.6m (339 µm, *Figure 5.7B*), where the grainsize increases from a fine silt/clay (*Figure 5.6* and *Figure 5.7A*) to a medium/coarse-grain sand, before once again returning to a fine silt/clay by 0.7m, whilst the mean grainsize for the upper 1m is 59 µm (coarse silt) (*Figure 5.6*).

Below 1m, the sediment suddenly changes from organic peat into an increasingly minerogenic, deep orange/brown silty/sand (grainsize =7.51 µm at 1.1m depth) (*Figure 5.6*). These pallid clays (mean grainsize = 50 µm) are interbedded with darker more organic sections until 1.4m (*Figure 5.9*), where the darker features become less prominent, and the sediment colour becomes increasingly pallid and sandy below 1.4m (*Figure 5.7C* and *Figure 5.9*). Below 1.5m palaeo-roots can be seen, creating a high cohesiveness, forming a blocky sediment structure between 1.56-1.65m (*Figure 5.9*), before becoming increasingly sandy below 1.7-2m (*Figure 5.7b*).
Below 1.8m depth (grainsize = 136 µm) the sediment becomes mostly sand (mean =63%, max =76%), whilst the grainsize shows high variability however predominantly remaining a grainsize of 129 µm until 2.6m (grainsize = 197 µm) (*Figure 5.6*). The remains of palaeo-roots can be seen again between 1.85-2.2m and 2.35-2.6m as manganese root casts, where the texture becomes blocky, condensing into large clay blocks. Between the two sections of root casts, the sediment shows evidence of some light iron mottling at 2.1-2.2m depth, shown in *Figure 5.10*.

![Figure 5.10: Core photograph for sedimentary structure at 1.9-2.28m (top) and 2.25-2.65m (bottom).](image)

Below the detrital minerogenic sediments at 2.67m, there is a sharp boundary from silty-sands into a highly organic, peat (*Figure 5.11*). This unit is 11cm thick, and has been dated using AMS C¹⁴ dating techniques to be between 35,090 cal. years BP (*Table 4.1*). This charcoal lens has a significantly smaller grainsize (4.4 µm, *Figure 5.3D*) than the minerogenic sediments above (197 µm at 2.6m) with a sub-angular, blocky organic sedimentary texture (1-3 mm width).

![Figure 5.11: Core photograph for sedimentary structure at 2.6-2.9m.](image)

Below the charcoal lens, there is a small slightly richer cream/brown coloured silt lens which shows good cohesiveness (*Figure 5.11 and Figure 5.12*). The grainsize = 197 µm (fine-grain sand), which is described as very poorly sorted (*Figure 5.6*). Below the sandy/silt lens, there is another dark (black), highly organic peat layer, with sub-angular blocky charcoal fragments (0.5-1 mm width), 9cm deep (*Figure 5.12*). The mean grainsize for this charcoal lens is 4 µm, very similar to the charcoal layer 2.67-2.78m depth.
Below 2.96m, the grainsize increases into a fine-grain sand (Figure 5.12 and Figure 5.7E) (grainsize = 275 µm) before there is a coarse grainsize horizon at 3.15m (grainsize = 742 µm) reaching maximum grainsize (Figure 5.6). Below 3.15m, there is a dark clay plug, approximately 15cm thick (Figure 5.12).

*Figure 5.12: Core photograph for sedimentary structure at 2.7-3.3m.*
5.1.1.3 LB-LN HA1 – Lake Baraba/Lake Nerrigorang sill core

HA1 is 5.8m deep auger hole conducted on top of the sill between Lakes Baraba and Lake Nerrigorang (Figure 4.1), consisting of predominantly sand (Figure 5.13).

The grainsize in HA1 varies between 38-532 µm, classified as coarse silt to medium grained sand following a general increasing grainsize with depth (Figure 5.13). There is a significant grainsize increase at 1.25m depth (grainsize = 437 µm, Figure 5.14b) and 4.5m depth (grainsize = 473 µm). The main constituent of HA1 is clearly sand, falling below 50% sand only at 4.1m depth.

![Figure 5.13: (Left to right) Sediment type (blue = sand%, red = silt% & grey = clay%) for LBLNHA1.](image-url)
Figure 5.14: Grainsize graphs from LB-LNHA1 from the Malvern Mastersizer 2000. (From top to bottom) A = 0.5m, B = 1.25m, C = 3.2m & D = 5m depth.
The upper 1.1m of LB-LNHA1 is composed of dark poorly sorted fine-grain sand (mean = 201 µm) (*Figure 5.13 and Figure 5.14A*). At 1m depth, the soil becomes more cohesive, forming small 0.5-1cm sediment clasts (*Figure 5.15A*).

At 1.25m depth, there is a significant increase in grainsize (437 µm) representing a poorly sorted, fine-medium grain sand. Below 1.25m depth (*Figure 5.14B*), the grainsize shows a distinct gradual increase until 2m, where the sediment becomes a very poorly sorted clean and pallid fine grain sand (*Figure 5.13 and Figure 5.15c*).

Below 2m depth, the grainsize increases until 2.5m (213 µm) where the colour is a pallid brown colour. Below 2.5m, the grainsize begins to decrease again, showing a gradual decrease until 4m (104 µm) (*Figure 5.13*). The texture of the sediment shows increasing cohesiveness between 3-3.2m, condensing into small cohesive sedimentary clumps (1-15mm in diameter) (*Figure 5.14B*). Below 4m, the sediment becomes dark in colour (samples from 4.1m and 4.4m depth, *Figure 5.15C*).

The bottom 3m shows a sharp grainsize increase into a medium-grain sand (mean grainsize =447 µm) which displays increasingly poor sorting and leptokurtic distribution with depth until 5.8m (*Figure 5.14D*). The colour of the sediment varies between a light-quartz rich washed sand (4.5, 4.7, 5 and 5.3m depth), into a dark organic, sandy sediment (5.1m, 5.5m and 5.8m depth).
Figure 5.15: The major panel (C) shows a photograph of the 25 samples taken from HA1, which were used for grainsize analysis. The bags are placed in 1m depth profiles, with each row representing the following metre below the surface (increasing depth from left to right in each row). The top two panels A(0.5-1m) and B(2.5-3.2m) show the increasing cohesiveness of these corresponding layers of sediment.
5.1.2 Secondary Cores

5.1.2.1 BAR04 – Lake Baraba northern-margin/sill core
BAR04 is a 0.77m core, taken from the northern lake margin (3m from the water’s edge) of Lake Baraba (Figure 4.1). The upper 0.3m consists of semi-decomposed organic material (peat), which becomes increasingly humified with depth, forming small cohesive structures containing charcoal fragment. Below 0.3m, there is a small pallid clay lens (0.3-0.33m) which lies above a dark brown, coarse sand lens at 0.34m (Figure 5.16). Below this, the sediment fines into organic clay, decreasing in organic content whilst increasing its cohesiveness with depth.

![Figure 5.16: BAR04 core photograph.](image)

5.1.2.2 BAR05 – Lake Baraba northern-margin core
BAR05 is a 1.07m core taken from the northern lake margin (at the water’s edge) of Lake Baraba (Figure 4.1). The upper 0.35m consists of peat, which condenses into small cohesive blocks below 0.25m (Figure 5.17). Below 0.35m, the sediment becomes grades into dark-organic clay with a high cohesiveness becoming increasingly pallid with depth (until 0.7m). Below 0.7m, the colour begins to darken, becoming increasingly cohesive and fine grained with depth. Below 0.8m, black organic clasts become visible (30%). Beneath this organic clay, there is a sudden boundary to a light yellow/cream clay unit below 1.05m (Figure 5.17).

![Figure 5.17: BAR05 core photograph.](image)
5.1.2.3 LB-LN HA2 – Lake Baraba/Nerrigorang sill

HA2 is a 2.2m auger hole which was taken on the sill between lakes Baraba and Lake Nerrigorang (Figure 4.1) comprised of almost entirely sand (Figure 5.18). The grainsize varies between fine-medium grain sand (mean grainsize =103-417 µm).

The upper 1.4m of LB-LNHA2 displays a general coarsening of grainsize with depth (175-417 µm). The sorting within LB-LNHA2 follows a similar pattern in the upper 1.4m, working from poorly-to-moderately sorted by 1.4m displaying increasingly normal skewness. Below 1.4m, the grainsize shows large variation between fine-medium grained sand (grainsize varies between 103-369 µm).

Figure 5.18: (left to right) Sediment composition (blue = sand, red = silt, grey = clay) and Grainsize (mean µm) for LB-LNHA2.
5.2 $^{14}$C Radiocarbon Dating
A total of four samples were dated using radio AMS C$^{14}$ radiocarbon dating from cores BAR02 and BAR03. In BAR02 (from the centre of the lake), samples were dated from two depths (1.42-1.46m and 2.85-2.87m). The dated material consisted of plant fragments and soil/bulk organic content, respectively. BAR03 (modern lake margin), two sample were also dated from two depths (2.775m and 2.72-2.78m). All 4 four radiocarbon ages have been using a high probability density range method (HPD) and have been calibrated using the SHCal134 atmospheric calibration curve, the findings are summarised in Table 5.1.

Table 5.1: Summary of the AMS radiocarbon results for Lake Baraba. All samples were dated using the Accelerator Mass Spectrometer (AMS) standard delivery and calibrated using the SHCal13 calibration curve.

<table>
<thead>
<tr>
<th>Location</th>
<th>Core</th>
<th>Core Depth (m)</th>
<th>Dating Material</th>
<th>Conventional $^{14}$C Age (BP)</th>
<th>Age (cal. BP)</th>
<th>Probability</th>
<th>pMC</th>
<th>Lab Code</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lake Baraba</td>
<td>BAR02</td>
<td>1.42-1.46</td>
<td>Plant</td>
<td>3012 +/- 20</td>
<td>3012-3230</td>
<td>95.40%</td>
<td>68.7</td>
<td>Wk-47794</td>
</tr>
<tr>
<td>centre</td>
<td>BAR02</td>
<td>2.85-2.87</td>
<td>Soil/bulk organic content</td>
<td>9311 +/- 19</td>
<td>9311-10360</td>
<td>95.40%</td>
<td>31.4</td>
<td>Wk-47795</td>
</tr>
<tr>
<td>Lake Baraba</td>
<td>BAR03</td>
<td>2.775</td>
<td>Charcoal</td>
<td>29732 +/- 184</td>
<td>29732-34200</td>
<td>95.40%</td>
<td>2.5</td>
<td>Wk-4779</td>
</tr>
<tr>
<td>margin</td>
<td>BAR03</td>
<td>2.72-2.78</td>
<td>Charcoal</td>
<td>35089 +/- 353</td>
<td>35089-40450</td>
<td>94.40%</td>
<td>1.3</td>
<td>Wk-47796</td>
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</tr>
</tbody>
</table>
5.3 Organics analysis - %C, %N, and C: N ratio

We only have %C and %N data for core BAR03 (lake margin) due to technical difficulties with the Elemental Analyser, meaning we have been unable to create C: N ratio for the lakes centre. %TC values obtained from the Elemental Analyser have been interpreted to represent %TOC, given the lack of carbonates demonstrated by Barber (2018) and no reaction was observed when the ground sediments treated with HCl.

In BAR03, the %C and %N were measured in an Elemental Analyser to determine the %TOC and %N of at 0.1m increments, from which the C/N ratio was determined for BAR03. The %TOC value within BAR03 varies between 0-37% (mean = 11%), whilst %N varies between 0-1.86% (mean = 0.4%), creating a C:N ratio ranging between 0-355 (mean = 32). Both the %TOC and %N are largest in the upper 1m of BAR03 (Figure 5.19). Below 1m, the %TOC and %N drastically drop to <1.22%TOC and <0.1%N until 2.7m, where both %TOC and %N become more variable. There are two small spikes recorded in both the %TOC and %N at 2.7 and 2.9m respectively, the first (2.7m) %TOC = % and %N = 0.25%, whilst the second (2.9m) had 10.5 %TOC and 0.31%N.

The C:N ratio generated from BAR03 shows a slight increase throughout the upper 1m of BAR03, showing an increase again in the two pre Holocene peat layers (2.7m and 2.9m depth). The large spike seen in the C: N ratio at 0.6m depth is likely exaggerated as this %N value was likely created, by depletion in N at 0.6m, which has resulted in an unreadable sample to be attained by the Elemental Analyser. Below 1m, the C: N ratio peaks at 1.1m (C: N = 111), before a second and third peak at 2.7 (C: N = 32.5) and 2.9m (34.4), reflecting the enrichment seen in both the %TOC and %N (Figure 5.19).
5.4 Lake Baraba sediment geochemistry

Elemental counts were analysed at a 500μm resolution for cores BAR02 (Figure 5.20) and BAR03 (Figure 5.21). This method of ITRAX core analysis provides continuous in-situ elemental count data throughout the entirety of the core (units created by the ITRAX are elemental counts per second (cps)). The results of this ITRAX analysis are presented in the following sections below.

5.4.1 The behaviour of elements reflecting sedimentary, hydrological and chemical conditions of Lake Baraba

The elements Si, Al, Ti, K, Rb and Zr are commonly used to determine the influx of detrital material (Davies et al., 2015). In the centre of Lake Baraba, detrital minerals (Si, Al, Ti and K) show low counts and play very little role from 0-2.6m, before showing a small enrichment just below 2.7m. Below 2.75m, detrital mineral counts increase dramatically peaking at depths 2.8m and 3.1m (Figure 5.22).

On the modern lake margin of Lake Baraba, detrital minerals (Si, Al, Ti and K) show low counts in the upper 1m of BAR03, apart from a large singular peak at 0.34m which is recorded throughout all detrital minerals (Figure 5.22). Below 1m, the detrital minerals show a sharp increase in cps, however remaining fairly uniform until 2.9m. Si and Al show enrichment between 2.6-2.85m (829-1551 and 64-143 cps respectively) whilst Ti and K show increased enrichment between 1.12-1.74m (ranging from 3374-11778 and 277-1750 cps respectively), becoming increasingly variable below 1.8m. Below 2.6m the detrital minerals become erratic, displaying two periods of enrichment (2.7m and 2.86-2.96m) and two periods of depletion (2.65-2.66m and 2.99-3m), before once again becoming enriched at 3.05m, the base of BAR03.

The Inc/Coh ratio and Br (cps) can be used to infer the amount of biogenic carbon and Br production by living aquatic organisms living within a lacustrine system (Phedorin et al., 2008; Fedotov et al., 2012). Further, low correlation values between biogenically produced elements such as Br and the Inc/Coh ratio compared to other detrital minerals.

In the centre of Lake Baraba, Br shows steady accumulation until 2.35m, before becoming enriched until 2.85m (9,310 cal. years BP, table 5.1) where the Br cps reach 1114 cps, before dropping off to <119 cps below 2.9m (Figure 5.21). Similarly, the Inc/Coh represents high stable values (6-8) throughout the upper 2.85m before a slight section of variable decline before settling around an Inc/Coh ratio of 5 below 2.19m depth (Figure 5.21). Meanwhile, on the Lake margin Br follows a similar pattern showing enrichment throughout 0-1m depth at 2.7m and 2.9m depth in the organic peat layers described in Chapter 5.1.1.2. The Inc/Coh ratio is elevated in the upper 1m (Inc/Coh =
6.5-7.5 cps) before sharply decreasing showing variability between a ratio of 4-5, before reflecting an increased ratio in the organic layers described at 2.7m and 2.9m depth (Figure 5.21).

The mobility of Fe and Mn creates a ratio which is known to be redox driven, and can therefore be used to infer the past redox conditions of a lacustrine environment (Fedotov et al., 2012; Davies et al., 2015; Davison, 1993). During anaerobic lake conditions (i.e. at the sediment water interface), Mn becomes significantly reduced and released into the water column causing a direct enrichment of Fe in the sediment (Davies et al., 2015). Therefore, increased Fe/Mn ratios are inferred to indicate increasingly anaerobic conditions at the sediment-water interface.

In the lake centre, Fe/Mn shows a steady decrease through the entire core, whilst the Fe/Ti ratio shows different sections of enrichment in the upper 0.3m and between 1.85-2.9m depth (Figure 5.20). Meanwhile on the lake margin, Fe/Mn shows a decreasing trend in the upper 1m of sediments, followed by a period of light enrichment until 1.45m and again at 2.7m and 2.9m in the organic (peat) layers described in 5.1.1.2 (Figure 5.21). The Fe/Ti ratio shows a slightly different trend, becoming enriched between 0.6-1m depth, before returning to baseline showing enrichment again at 2.7m and 2.9m depth.

The palaeo-environmental indicators Ca/Ti and Sr/Ti indicate increased Ca and Sr precipitation in a lacustrine environment, produced via biogenic precipitation (Moreno et al., 2007; Litt et al., 2009). In Lake Baraba, Ca/Ti and Sr/Ti show very similar relationships throughout each core on both the Lake margin and in its centre (Figures 5.20 and 5.21). In the centre of the lake (BAR02), Ca/Ti and Sr/Ti show enrichment in the upper 0.3m, before showing low values throughout the rest of the core (Figure 5.20). On the margin of Lake Baraba, the Ca/Ti and Sr/Ti show enrichment in the upper 0.2m before relatively low values until before becoming enriched again between 0.6-1m depth. Similar to the lake centre, both the Ca/Ti and Sr/Ti ratios become consistently negligible below the organic peat units (upper 1m of BAR03) until 2 small peaks within lower the organic layers (2.7 and 2.9m, BAR03) (Figure 5.21).

In BAR02, Si shows a high correlation with all elements, particularly K ($r^2 = 0.99$) and Ti ($r^2 = 0.98$), whilst also showing strong correlation with Rb ($r^2 = 0.92$), Sr ($r^2 = 0.96$) and Zr ($r^2 = 0.94$). Other elements such as Ti show strong correlations with Rb ($r^2 = 0.97$), Sr ($r^2 = 0.97$), Zr ($r^2 = 0.95$) and K ($r^2 = 0.99$), whilst other notable mentions include Sr-Zr ($r^2 = 0.97$), Sr-K ($r^2 = 0.97$) and Fe-Mn ($r^2 = 0.96$) (Table 5.2)

In BAR03, the detrital minerals show significantly less correlation than seen in BAR03. Si demonstrates a high correlation with K ($r^2 = 0.80$), Ti ($r^2 = 0.78$), Al ($r^2 = 0.82$) and Zr ($r^2 = 0.83$).
Meanwhile, Ti shows correlation with Rb \( (r^2 = 0.76) \), Zr \( (r^2 = 0.94) \), Al \( (r^2 = 0.88) \), K \( (r^2 = 0.82) \) and Ca \( (r^2 = 0.76) \), whilst Al has high correlation values with K \( (r^2 = 0.88) \) and Zr \( (r^2 = 0.83) \) (Table 5.3).

**Table 5.2: BAR02 geochemical correlation values**

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<tr>
<th></th>
<th>Al</th>
<th>Si</th>
<th>K</th>
<th>Ca</th>
<th>Ti</th>
<th>Mn</th>
<th>Fe</th>
<th>Rb</th>
<th>Sr</th>
<th>Zr</th>
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<td></td>
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<tr>
<td>K</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Ca</td>
<td>0.657789</td>
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<tr>
<td>Ti</td>
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<tr>
<td>Mn</td>
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<td>0.212399</td>
<td>0.380608</td>
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<td>Fe</td>
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<td>0.359925</td>
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<td>0.367895</td>
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<td>Rb</td>
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<td>0.460097</td>
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<td>0.497111</td>
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<td>Zr</td>
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<td>0.810408</td>
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**Table 5.3: BAR03 geochemical correlation values**

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<td>K</td>
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<td>Ca</td>
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<td>0.912944</td>
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<td>0.904568</td>
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</table>
Figure 5.20: BAR02 lake productivity, Redox and lake level/salinity proxies.

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Figure 5.21: BAR03 lake productivity, Redox and lake level/salinity proxies.
Figure 5.22: BAR02 and BAR03 Detrital minerals (Al, Si, K, Ti and Rb).

Catchment Erosion Proxies

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5.5 Summary of results

The different control mechanisms (chronology, facies and geochemical analysis) used to recreate the palaeo-environments of Lake Baraba show demonstrate different palaeo-environments evidently recorded in the sediments of Lake Baraba. The modern peat production and accumulation in Lake Baraba was found to initiate ~9.3ka cal. years BP, whilst pre-Holocene peat forming environments were discovered at the lake margin at 2.7m and 2.9m (~35ka cal. years BP). In between the zones of peat formation within Lake Baraba, a series of oxidised minerogenic sediments are visible in which it is obvious that organic accumulation was not favoured, indicating a different sedimentary environment through this time in Lake Baraba’s palaeo-history.
6. DISCUSSION

Analysis of the stratigraphy and elemental composition of the sedimentological data of Lake Baraba implies that Lake Baraba has undergone significant hydrological change in the past. This chapter will explore these results, firstly at their face value discussing the C: N content of the lake margin, chronology, stratigraphic and elemental relationships within Lake Baraba. Collectively this information will be compiled to discuss sedimentological facies interpretation and peat forming environments in Lake Baraba, before relating this to palaeo-hydroclimate of Lake Baraba and its inter lake connectivity. The results gathered from Lake Baraba will then be discussed in the broad context of south eastern Australia.

6.1 Chronology and Stratigraphy

The cores taken along the LB-LN transect (Figure 4.1) contain relatively complex stratigraphy often displaying large shifts in grainsize, organic content, colour and texture becoming increasingly organic towards the centre of Lake Baraba. The stratigraphy of the LB-LN sill is unsurprisingly dominated by sandy deposits, which become increasingly diminished towards the centre of Lake Baraba; where peat and organic silts become the dominant lithology (Figure 6.2 and 6.3). The margin of Lake Baraba contains significantly less organic material than the Lake Couridjah margin described by Barber (2018) and is predominantly composed of dense clays below 1m (Figure 5.6). This section below discusses the details of interpretation of the stratigraphy presented in the results section, whilst the nomenclature provided for each facies in Figures 6.2 and 6.3 (e.g. L1a) will be used to refer to the units in the subsequent descriptions.

6.1.1 Sill and Lake Stratigraphy

The LB-LN sill is predominantly composed of fine to medium grain sands which are interbedded with fine-to-coarse grain sandy deposits pinching out towards Lake Baraba. As you move towards the lake centre, the sands are replaced by fine-grained lacustrine sediments before grading into heavily organic facies towards the lake centre.

6.1.2 Chrono-stratigraphy

To determine the chronostratigraphic relationship of The sediments in Lake Baraba, 4 organic carbon samples were selected from different depths within the cores BAR02 (1.42-1.46m (3,010 cal. years BP) and 2.85-2.87 (9310 cal. years BP)) and BAR03 (2.72-2.78m (35,090 cal. years BP)), the results of which are presented in Table 5.1. The chronology determined for Lake Baraba correlates with existing published chronologies from the TLNP (Barber, 2018 and Gergis, 2000) whom both dated the onset of L1 (Holocene peat facies) to be between 10,690-10,390 cal. years BP in Lake Couridjah,
whilst Black et al. (2006) estimated the start of peat accumulation in the centre of Lake Baraba to begin slightly later ~ 8.5 cal. years BP at 4m depth.

**Age/depth model**

From the AMS radiocarbon dates obtained from Lake Baraba, the chronology of stratigraphic units in the centre of Lake Baraba could be generated from the two samples in BAR02 (Table 5.1), from depths 1.42-1.46m and 2.85-2.87m (Figure 6.1). A similar model was not able to be determined for the lake margin, however dating the organic facies at 2.7m to be 35,090 cal. years BP (table 5.1). Hence, it can be inferred that the onset of current peat forming conditions in the centre of Lake Baraba commenced approx. 9,310 cal. years BP, whilst the bottom of BAR03 on the lakes margin at 3.23m depth could be >43,000 cal. years BP. The age/depth model created from Lake Baraba shows a similar pattern to the age/depth model created by Black et al (2006), whom reported the onset of peat forming conditions at 4m depth (~8.5 cal. years BP), whilst Barber (2018) found the onset of peat conditions to occur ~10.8 ka at 1.69m depth in Lake Couridjah.

*Figure 6.1: Age-Depth relationship from BAR02 (centre of Lake Baraba), calculated using two radiocarbon dated samples from depths 1.42-1.46m (3012yr BP) and 2.85-2.87m (9311 yr BP) (Table 5.1).*

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6.2 Geochemical Conditions and Sedimentary History of Lake Baraba

6.2.1 Changing hydrological and chemical conditions inferred from elemental data

Elements such as Ti do not feature within biological processes and are mainly found contained within detrital sediments containing weathering resistant minerals such as rutile (TiO₂). This allows Ti to be an extremely useful baseline proxy to which temporal changes in other detrital and biogenic minerals can be measured (Davies et al., 2015; Kylander et al., 2013). Secondly, given that the TLNP is surrounded by a quartz-rich Hawkesbury Sandstone, Si can also be considered as an appropriate measure of detrital material influx to Lake Baraba from the surrounding catchment in the form of quartz (SiO₂) (Barber, 2018). However, caution should be taken when adapting Si for sediment delivery as Si can also be produced via authigenic biogenic Si production (BSi) (Davies et al., 2015).

Delivery of sediment

On the lake margin, the detrital minerals (Al, Si, Ti and K) show little activity in the upper 1m (Holocene peats); however experience a small peak just below 0.3m depth (Figure 5.22). High correlation values between Ti and Si (r = 0.98) in the lake centre, indicates similar influx patterns in delivery of both Ti and Si from the surrounding catchment (Davies et al., 2015), therefore signifying a low abundance of BSi producing organisms within the centre of Lake Baraba. On the margin of Lake Baraba however, lower correlation values (r = 0.77) between Ti and Si either indicate a) a higher BSi production than the centre of Lake Baraba or b) a different pattern of sediment delivery from the surrounding catchment (Davies et al., 2015). To a lesser extent, the high correlation values between Si, Ti, K and Rb (Table 5.2b) indicates that the detrital material entering Lake Baraba contains large amounts of weathering resistant silicates (particularly K-feldspars and Micas) involved in the sedimentary influx to Lake Baraba (Tallon-Armada et al., 2014).

Lake productivity

The Inc/Coh ratio measured during core irradiation by the ITRAX core scanner indicates the mean number of scattering between elements within the sediments, in which the number of incoherent scattering between electrons is higher for elements with a low atomic mass (i.e. high carbon content). Therefore samples with a higher organic content will produce a larger Inc/Coh carbon ratio (Guyard et al., 2007; Roberts et al., 2016; Jenkins, 1999). Verification of this is provided, as the Inc/Coh ratio produced by the ITRAX reflects the result of %TOC from the elemental analysis of BAR03 (Figure 5.21), whilst Barber (2018) also found the Inc/Coh atomic scattering ratio to effectively mirror the %TOC of Lake Couridjah. Therefore %TOC from Lake Baraba will be derived from the Inc/Coh scattering ratio and %C for BAR03 as no carbonates were found in the sediments of Lake Baraba (Davies et al., 2015).
In both the centre and lake margin of Lake Baraba, the Br count highlights areas of increased enrichment in the upper and lower sections within the Holocene peats, possibly indicating different water depths in Lake Baraba throughout the Holocene record. Below the Holocene peats, the Lake productivity proxies show very little until showing small periods of enrichment in at depths of 2.7m and 2.9m, coinciding with the pre-Holocene peats dated to be 35,090 cal. years BP (table 5.1, Figures 5.20 and 5.21).

**Redox conditions**

The mobility of Fe and Mn has a known redox driven ratio and can therefore be used to infer the past redox conditions of a lacustrine environment (Fedotov et al., 2012; Davies et al., 2015; Davison, 1993). During anaerobic lake conditions (i.e. at the sediment water interface), Mn becomes significantly reduced and released into the water column, causing a direct enrichment of Fe in the sediment. Therefore an increase in the Fe/Mn ratio can be interpreted to indicate increasingly anaerobic conditions in Lake Baraba (Davies et al., 2015). Such changes in redox conditions of Lake Baraba could be caused by one of two environmental changes; a) increased stratification in Lake Baraba caused by a deeper water level, or b) decrease in wind strength and frequency in the TLNP which has resulted in a lesser degree on lake water mixing (Moreno et al., 2012; Davies et al., 2015).

In the lake centre, Fe/Mn ratio indicates increasingly aerobic conditions throughout the Holocene, whilst the Fe/Ti ratio shows different sections of enrichment in the upper 0.3m and between 1.85-2.9m depth (Figure 5.20). On the lake margin, the Fe/Mn ratio shows a steady decrease throughout the Holocene peats. The Fe/Ti ratio on the other hand, shows a different pattern of redox condition in Lake Baraba (Figure 5.20 and 5.21). The Fe/Ti ratio indicates increased reducing conditions in the upper section of the Holocene peats, before a period of relatively well oxygenated water column during the mid-Holocene before anoxic conditions becoming dominant again throughout the early-Holocene peat accumulation (Figures 5.20 and 5.21). Below the Holocene peats, oxidising conditions appear to once again become dominant, represented by a drop in Fe/Ti and Fe/Mn below 2.9m (BAR02), whilst on the lake margin the sediment colour become yellow/orange/brown indicating highly oxidated water column. Redox mineral ratios Fe/Mn and Fe/Ti become increasingly invariable with depth until the organic zones BAR03 on the lake margin (Figures 5.20 and 5.21) in the late-Holocene peat formations.

**Lake water level/salinity**

The palaeo-environmental indicators Ca/Ti and Sr/Ti indicate increased Ca and Sr values (usually in the form of carbonates such as CaCO$_3$ and SrCO$_3$, however the presence of carbonates was not found in the sediments of Lake Baraba so these elements must be hosted by different minerals) in
the upper zone 1 indicating a high level of Ca and Sr precipitation during the late-Holocene peats in the centre of Lake Baraba (Moreno et al., 2007; Litt et al., 2009). Below, Ca/Ti and Sr/Ti ratios indicate very little to no Ca and Sr evaporative precipitates in the centre of Lake Baraba (Figure 5.20). On the lake margin however, the Ca/Ti and Sr/Ti ratios indicate enrichment in the late-Holocene peats, whilst relative stability through mid-Holocene before once again showing significant enrichment throughout the early-Holocene (Figure 5.21). The different levels of Ca and Sr precipitation between the lake centre and its margin throughout zone 3 could indicate; a) increased mineral precipitation at the lake margin than its centre as a direct response to changing water depths which is recorded at the margin much better than the centre of the lake or b) erosion, evaporation and palaeo-environmental proxies could receive some protection and/or insulation from the floating peat islands in the centre of the lake, as described by Vorst (1974). Meanwhile, the Sr/Ca ratio highlights a different trend throughout the Holocene peat development in Lake Baraba, and is interpreted to indicate increased in-situ Sr precipitation due to increased evaporative Ca concentration within the lake waters (Kylander et al., 2011; and Martin-Peurtas et al., 2009).
6.3 Facies interpretation

In the following section the sill and lacustrine environments will be divided into their individual facies with their paleo-environmental context and meaning to be discussed. Broadly, there are two primary units of sediment apparent along transect LB-LN transect; termed the Primary unit 1, consisting of four different sill facies, and Primary Unit 2, which consists of 5 different lacustrine facies. Each facies was discerned through a cluster analysis (Figure 6.2 and Figure 6.3) and the subsequent facies analysis will be described sequentially below.

6.3.1 Primary Unit 1: Sill deposits

Sill facies 1 (S1)
Sill facies 1 is a dark, fine grain silty/sandy alluvium which forms small clasts (0-2.5cm, Figure 5.15a), apparent at shallow depths 0-1.1m in the sill, becoming interbedded with facies S2 and S3 (between 0.7-2m, Figure 6.2). S1 once again appears between 2.5-3.5m in HA1, forming small clasts 0.5-1.5cm in diameter (0.5-1.5cm, Figure 5.15b and 5.15c). The dark organic colour and increased silt content (13-33%) and clay content (11-21%) of facies S1 infer the presence of palaeo-soil development within the LB-LN sill. The implication of palaeo-soils in the sill formation infers that at the time of deposition, Lake Baraba was; 1) a wetter catchment and Lake Baraba’s water levels were high, 2) increased moisture content within the sill (i.e. impeded drainage system) and 3) a relative geomorphic stability within the TLNP catchment, as palaeo-soils have been inferred to suggest the sedimentation delivery rate is slower than the rate of soil formation on the sill (Zhou, 1994; Barber, 2018).

Sill facies 2 (S2)
Sill facies 2 consists of poorly to very poorly sorted-fine grain sand, comprising of a very high sand content (88-90%) whilst containing very little silt (8-13%) and clay (2-9%) components (Figure 6.2). S2 appears at higher depths in the LB-LN sill, interpreted as the physical barrier that separates Lakes Baraba and Nerrigorang. Facies S2 was likely deposited by small ephemeral streams operating on the alluvial fan extending from the adjacent from the sandstone ridgeline (Russel et al., 2010).

Sill facies 3 (S3)
Sill facies 3 is found at deeper depths in the LB-LN sill, apparent in both cores HA1 (1.2m and below 4.5m) and HA2 (1-1.6m and below 2m). S3 consists of poor to moderately sorted, coarse grain sand (92-95% sand composition, Figure 6.2). Similarly to S2, this sand is thought to be deposited through fluvial processes’ carrying sediments from the Hawkesbury sandstone ridgeline, along the alluvial fan and depositing them in the LB-LN sill(Figure 6.2).
Figure 6.2: Sill facies for HA1 (a), HA2 (b) and the cluster analysis for the sill facies determination (wards method). The accompanying principle components analysis (PCA) is included in APPENDIX B.

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6.3.2 Primary Unit 2: Lacustrine deposits

The lacustrine facies found in Lake Baraba consist of two distinct groups; 1) organic peats and peaty-clay units found in the lake centre and the upper 1.4m of the current lake margin, and 2) minerogenic units, found in the lake centre (below 3m) and on its margin (below 1m), distinguished by a greatly enlarged grainsize and increased variability of sorting values (highlight greatly increased grainsize variable sorting (Figure 6.2 and 6.3) (PCA analysis attached to Appendix b).

6.3.2.1 Organic (peat) units (BAR02 and BAR03)

The dominant organic units (facies L1, peat and peaty clays) which are found in Lake Baraba can be divided into 4 distinct facies (Figure 5.20 and 5.21), based upon their grainsize (<33 µm), sediment composition (high silt and clay content) and sorting to a lesser degree (Figure 6.3).

Lacustrine facies 1a (L1a)

Lacustrine facies 1a is described as poorly-to-very poorly sorted clay (grainsize ranging between 5-11 µm). L1a is apparent below 0.5m becoming interbedded with L1b below 1.2m in the lake centre (BAR02), pinching to only 1.1m depth at the lake margin (BAR03, Bar04 and BAR05). L1a once again becomes apparent at 3.23m depth (BAR03), rising right back up to near the surface (0.35m depth) in BAR04 and BAR05 on the northern margin of Lake Baraba (Figure 5.17, 5.18 and 6.3). The sediment composition of L1a consists of mostly silt (58-73%) and clay (25-37%), containing a small sand component (<8%) sand. L1a coincides with periods of significant Ca and Sr precipitation, associated with increased anoxic conditions. Enrichment of Br associated with the late-Holocene peat development associated with L1a, could infers a decreasing lake water level, evident in the lake centre and at its margin.

Lacustrine facies 1b (L1b)

Lacustrine facies 1b is comprised of very poorly sorted very-fine silty/clay. L1b has a slightly larger mean grainsize than the other organic facies L1a and L1c (Figure 6.3), caused by slightly enriched sand content (15-33%) at the expense of clay content (15-33%) (Figure 6.3). Facies L1b occurs in the upper 0.5m of BAR02 (lake centre), appearing between 0.7-1.3m in BAR03 (western-lake margin) and above 0.35m depth in BAR04 and BAR05 (northern-lake margin). The occurrence of L1b at lower depths at the lake margin compared to the lake centre could indicate differing processes of gradual water level decrease on the lake margin recorded in the marginal depositional environment. Facies L1b represents the mid-Holocene peats which show a decrease in peat accumulation, which may also possibly infer a slightly higher lake level during the time of deposition.
**Lacustrine facies 1c (L1c)**

Lacustrine facies 1c (L1c) consists of moderate to poorly sorted clay, which has a slightly smaller mean grain size and better sorting co-efficient than L1a and L1b (*Figure 6.3*). L1c is apparent below 1.3m in the lake centre (BAR02), interbedded with L1a becoming more dominant with depth (*Figure 6.3*). At the lake margin, L1c can be seen at 2.7m and 2.9m depth, comprising of almost entirely silt (58-73%) and with a very high clay content (38-49%). Facies L1c is indicative of both Holocene and pre-Holocene peat development in Lake Baraba, possibly representing a similar facies L4 described by Barber (2018). Barber (2018) described the presence of dark-fine grain sediment interpreted to represent an extensive swampy phase, culminating in undisturbed swampy conditions connecting Lakes Baraba and Couridjah. The continuous swampy conditions between Lake Baraba and Couridjah would have only been enabled without the presence of a separating sill structure (L4 is described to occur 2-2.5m deep at the Lake Couridjah/Lake Baraba margin, Barber (2018)).

**Lacustrine facies 1d (L1d)**

Lacustrine facies 1d is a small facies found at the surface of BAR03, between 0.3-0.4m depths, consisting of almost entirely clay (94-100%) and a very small grain size (*Figure 6.3*). L1d shows close correlation to the organic facies L1a-c, however is displayed as its own distinct group consisting of almost entirely clay, with a very small grain size readout from the Malvern Mastersizer (<10 µm, *Figure 6.3*). A core photograph showing the high organic content of the upper 0.4m of BAR03 can be seen in *Figure 5.8*.

### 6.3.2.2 Minerogenic units (BAR03)

Lacustrine facies 2-5, contain significantly lower organic composition than the highly enriched (%TOC and Br (cps)) of the organic (peat) facies 1a-d above them (*Figure 6.3*).

**Lacustrine facies 2 and 4 (L2 and L4)**

Lacustrine facies 2 is consists of poorly to very poorly sorted-coarse silt, consisting of mostly sand (48-60%) and silt (31-41%). Facies L2 can be seen on the lake margin at 0.5m, before appearing again interbedded with L4 between 1.4-2.4m (BAR03, *Figure 6.3*). Catchment erosion proxies Rb, Ti, K and Si show increased abundance in facies L2, likely inferring either; a) a wetter catchment, b) less vegetated environment or c) increased high energy events washing larger grain size sediment into Lake Baraba.

Lacustrine facies 4 is a very fine grain sand (67-75% sand) first appearing at 1.8m depth. Below 1.8m, L4 becomes interbedded with facies L2 on the lake margin until 2.6m, before appearing again in between the two pre-Holocene peat facies at 2.7m and 2.9m (*Figure 6.3*). Facies L4, similar to L2
possibly represents deepening water levels in Lake Baraba, deposited during wetter conditions around the same time of the pre-Holocene peat formation (~35,090 cal. years BP).

**Lacustrine facies 3 and 5 (L3 and L5)**

Lacustrine facies 3 and 5 are described as a fine to medium grain sand, slighter larger in grain size than facies L2. Facies L3 and L5 consist almost entirely of sand (95%) on the lake margin, whilst facies L3 appears at two locations in BAR03 (0.6m and 3.1m depth, Figure 6.3). Facies L3 contains a high silt (14-16%) and clay (12%) content, whilst L5 contains almost exclusively sands likely associated with a drier, more arid environment in the catchment area.
Figure 6.3: (a) facies overlay for the lake centre (BAR02), (b) facies overlay for the lake margin (BAR03) and (c) lacustrine facies cluster analysis used to determine the lacustrine facies characteristics (Wards method). The accompanying principle components analysis (PCA) can be found in Appendix b.
6.5 Stratigraphy, groundwater and connectivity of Lake Baraba and other TLNP systems

The similarity in stratigraphic relationships inferred between sedimentary sequences of Lakes Baraba and Couridjah, such as facies L4 (highly organic peat facies found in Lake Baraba and Lake Couridjah, described by Barber (2018)), have also been found in Lake Baraba core BAR03 (2.7m and 2.9m (Figure 5.19)) dated to be 35,090 cal. years BP. The appearance of this continuous swampy facies in both Lakes Baraba and Couridjah (at similar depths and ages) implies the two lakes have experienced differences in connectivity throughout the late-Quaternary, before becoming completely separate lacustrine systems. During the time of the study, Lake Couridjah was experiencing dramatically declining water levels, recharging the subsurface groundwater below however Lake Baraba did not show the same symptoms of a falling water level. Therefore the retention of water in Lake Baraba compared to Lake Couridjah could be a process of either; a) Large floating peat islands on Lake Baraba which provide insolation preventing evaporation of the lake waters, or b) Lake Baraba contains an aquiclude at its base preventing interactions between surface and groundwater systems (Vorst, 1974; Gergis, 200; Black et al., 2006). It is likely a combination of these factors in Lake Baraba leading to a greater water level of water retention within Lake Baraba in comparison to other lakes in the TLNP.

Assuming the sub-surface interpretations of Lake Baraba are correct, there is potentially direct evidence found at 2.7m and 2.9m depth at the Lake Baraba margin (BAR03) indicating possible continuous swampy conditions which extended between Lake Baraba and Couridjah (and potentially Lake Nerrigorang) 35,090 cal. years BP (Figure 6.1), inferred to be >38,250 cal. years BP (at 2.4m) in Lake Couridjah, however this date was extrapolated not directly dated (Barber, 2018). It is likely that since this time, however; Lake Baraba and Couridjah have not retained a connected hydrology due to the presence of an aquiclude found to line the base of Lake Baraba in cores BAR03, BAR04 and BAR05. During times of a wetter climate however, lake water levels have the potential to rise and overflow the sill structures creating a larger interconnected system between Lake Baraba and Lakes Couridjah and Lake Nerrigorang (Barber, 2018; Gergis, 2000). Variations in the hydrological interconnectivity of Lake Baraba and its various features (floating islands and aquiclude) potentially allow Lake Baraba to retain water during drought conditions better than the other lakes in the TLNP.
6.6 Paleo-environments and their Context within South-east Australia

During the late-Quaternary period, south-east Australia experienced a variety of conditions ranging from relatively cool to intermediate temperatures, highlighted by measured changes in sea surface changes (Barrows et al., 2007), lake levels and vegetative communities and Aeolian depositary phases (Lomax et al., 2011).

With the lack of age control before 2.72m depth at the lake margin, it is difficult to infer the ages of the oxidised sediments in BAR03; however, a date was attained for the pre-Holocene peats at 2.72-2.78m (35,090 cal. years BP.). Therefore, core BAR03 indicates two layers of peat deposition in Lake Baraba occurring approximately 35,090 cal. years BP and 9,310 cal. years BP. Therefore BAR03 contains a potential climatic record dating to before the Early Last Glacial Maximum of the late Quaternary period (~35-22ka) (Petherick et al., 2013; Gergis, 200; Markgraf et al., 1992).

The early last glacial period (~35-22ka) in south-east Australia brings the (Marine Isotope Stage) MIS 3 into MIS 2, representing a time of moist climatic conditions producing high fluvial activity, observed to occur in many large south-east Australian rivers (van Meerbeeck et al., 2009; Petherick et al., 2013; Markgraf et al., 1992). Regionally significant rivers such as the Lachlan, Murrumbidgee, Murray and Goulbourn rivers all developed large paleo-channels during the early LGM (Petherick et al., 2013), while increased moisture in Lake Baraba is evident through the presence of the pre-Holocene peat layers (35,090 cal. years BP) found at 2.7m and 2.9m depth on the lake margin.

Following the early LGM, the climate of south-east Australia evolved into a drier, cooler climate in coincidence with the LGM (~22-18 ka). The LGM brought increased aridity, evapotranspiration and wind, whilst rainfall levels decreased to half the modern precipitation levels of south-east Australia (Petherick et al., 2013; Gergis, 2000). During the LGM, the landmass of Australia increased in size resulting in widespread drought and lake level decline across south-east Australia, evidenced by a mostly treeless landscape dominate by grass steppe above 600m AHD (Petherick et al., 2011; Kemp and Hope, 2014). These dry arid conditions were likely not experienced in the TLNP as Black et al. (2006) indicated by showing Casuarinaceae dominance in the pollen records of Lake Baraba, indicating swampy conditions in the Lake Baraba surrounds. The Casuarinaceae dominance highlights the TLNP may have acted as a form of refugium during the LGM, reflecting a swampy environment with moist conditions (Black et al., 2006). Increased oxidative conditions (Figure 5.21) are apparent during the LGM, inferring greater mixing of the lake waters and decreased stratification, likely caused by increased windiness throughout the LGM, whilst increased detrital inflow to the Lake Baraba (Figure 5.22) infers an increasingly arid landscape, leading to increased erosion from the catchment into Lake Baraba. Water levels during the LGM show no signs of complete drying.
throughout the LGM in our cores obtained, as found by Black et al. (2006) however lake productivity and peat formation and accumulation is not evident throughout the LGM (Figure 5.21). The decreased organic accumulation is possibly indicative of deeper water levels, where shallow water sedges were unable to grow, often growing in water that is only 2-4m deep (NPWS, 1995; Gergis, 2000; Fairley, 1978).

Following the LGM, the subsequent glacial-interglacial transition period (~18-12ka) is characterised by a rapid retreat of glaciers in south-east Australia (Barrows et al., 2001) coupled with the re-emergence of arboreal species throughout temperate south-eastern Australia (Williams et al., 2009; Petherick et al., 2013). The climate from south-east NSW in the Southern Highlands (29km to the south-west of the TLNP) remained relatively arid throughout the late LGM (Kemp and Hope, 2014) recorded by significant decreases in lake water levels in Lake George ~14-10ka (142km to the south-west) (Fitzsimmons and Barrows, 2010). During this period, Lake Baraba shows very similar conditions to those it experienced in during the LGM, however the Fe/Mn and Sr/Ca ratios indicate a transition towards more anoxic conditions. The increasing anoxic conditions possibly reflect either: a) decreasing water level of Lake Baraba throughout the interglacial period as it approaches the Holocene or, b) decreased windiness increasing the effects of temperature stratification on the waters of Lake Baraba. Lake water level proxies, as well as Lake productivity values show relatively little variability through the interglacial, indicating stable environmental conditions present in the TLNP, reflecting a general water level decline into the early-Holocene where peat production is reinitiated (Figure 5.21).

The onset of the Holocene (~12 ka) in south-east Australia is characterised by increased temperatures and precipitation levels with a generally stable climate (Kemp and Hope, 2014). The conducive climatic conditions encouraged the establishment of aquatic vegetation and preservation of organic sediments within lakes and swamps, seen in Lake Baraba (~9,310 cal. years BP) (Fryirs et al., 2014). These changes are recorded well within the sediments of Lake Baraba, with detrital input proxies showing significantly decreased influx during the Holocene, whilst increases in lake productivity indicate the transition period from a lacustrine environment to a peat forming depositional system at around between ~9310-10360 cal. years BP (Figures 5.20, 5.21 and 6.1). This increasingly moist climate is reflected throughout south-east Australia as Lake George experienced its highest Holocene shore levels 10-8 ka (Fitzsimons and Barrows, 2010), whilst Mountain Lagoon in the Blue Mountains persisted as a swamp before indicating lacustrine conditions ~10 ka (Rose and Martin, 2007). During the early Holocene, evidence from Lake Baraba indicates increased organic material production and its subsequent accumulation,
whilst water level proxies indicate high precipitation of Ca and Sr indicating a decreasing lake level. Increased anoxic conditions in the early-Holocene peats observed could be a product of increased organic material build up deposition covering both the surface and floor of Lake Baraba with organic material, reducing the oxygen amounts in the lake.

By the mid-Holocene (~6 ka), the climate of south-eastern Australia had returned to drier and cooler conditions displaying larger climatic variability, widespread throughout many publications (i.e. Dodson, 1987; Woodward et al., 2014; Harrison, 1993; Tyler et al., 2015). The increasingly variability of the climate throughout the late-Holocene is thought to be driven by the onset of El Nino Southern Oscillation (ENSO), with a dominant El Nino cycle producing dry climatic periods throughout the east coast of Australia (Reeves et al., 2013; Woodward et al., 2014).

Barber (2018) found evidence of drying periods in the mid-Holocene peat on the margin of Lake Couridjah between 0.33-0.35m depths, apparent by a small pallid sandy lens deposited approximately 4 ka; however this evidence is not seen on the margin of Lake Baraba. Black et al (2006) found a hiatus in peat deposition and increase in fungal spores between 6-5.2ka reflecting the increasing drying conditions of the mid-Holocene, however a lack of geochemical evidence in the sediments of Lake Baraba during the mid-Holocene infers a more stable and permanent water level in Lake Baraba than other lakes in the TLNP showing no evidence of complete drying periods during the Holocene unlike evidenced in Lake Couridjah (Barber, 2018).

Furthermore, evidence of climatic variability during the late-Holocene is apparent within the upper 0.2-0.3m of BAR02 in the lake centre, demonstrated by significant increases in Palaeo-environmental indicators indicating a decreasing lake level in Lake Baraba, which was also indicated to be the case in Lake Couridjah by Barber (2018). Barber (2018) found the presence of ‘crumbly’ peat (dating between ~1.8 – 2.2 ka) when lake evaporation started to increase in Lake Couridjah whilst lake productivity saw a slight decrease. Therefore, despite some evidence of hydrological variability in south-east Australia throughout the Holocene, the water level of Lake Baraba has remained relatively stable since ~9,310 cal. years BP, showing no evidence of complete drying throughout the Holocene, whilst Barber (2018) found Lake Couridjah had also maintained relatively stable water levels since ~10.8 ka however possibly experiencing a significantly decreased water level 4ka BP.

Therefore, evidence suggests that during the early last glacial period, Lakes Baraba and Couridjah were connected as one continuous swampy environment which became separated sometime during the onset of the LGM in south-east Australia. After a short period of dry, arid conditions during the deglacial period, Lake Baraba represented a small and shallow lake system with very little production.
or preservation of organic material throughout the glacial periods of the late-Quaternary. During this time between the LGM and the interglacial melting, the sill deposits separating Lake Nerrigorang, Baraba and Couridjah were formed isolating each lacustrine system before the onset of the Holocene (~12 ka) as observed today.

6.7 Implications of Results

Despite some evidence of hydrological variability throughout the Holocene, Lake Baraba has remained relatively stable and unchanged for the past ~9.3 ka cal. years BP (Lake Couridjah ~10.8ka (Barber, 2018)). Other south-east Australian lakes have experienced significantly large shifts from standing water conditions during the early-Holocene towards increased wetland environments throughout the mid-to-late-Holocene (i.e. Little Llangothlin Lagoon, (Woodard, 2017) and Mountain Lagoon, Blue Mountains (Robbie and Martin, 2007). Furthermore, pollen research by Black et al., (2006) suggests that the Thirlmere Lakes may have acted as a refugium for forest dwelling taxa during the LGM. All these findings indicate a relatively stable climate and environment present in the TLNP throughout the Holocene record (~12ka till present). Therefore further research is needed to understand the drivers of the significant modern changes in recorded in the lakes of the TLNP, not observed in the nearly complete Holocene peat record obtained from Lake Baraba (9.3 ka cal. years BP).

6.8 Limitations

Elemental analysis data

Elemental analysis data was not able to be attained in the centre of Lake Baraba (BAR02) due to ongoing technical issues with the elemental analyser, therefore meaning a reading for the samples prepared for BAR02 were unable to be attained for the purpose of this study.

BAR03 grainsize analysis

Grainsize data was unable to be gathered in the upper 0.3m of BAR03, due to the extremely high organic content of the peat samples leaving not enough sediment for analysis after the digestion process removing all the organic material. Therefore a significantly larger grainsize sample size is needed to be used to generate an accurate grainsize for this section of the core, or a different analysis or digestion method is required to generate an accurate and attainable grainsize reading for the upper 0.3m of the highly organic peat.
Age/Depth model only created for the lake centre (Chronology)
Due to the areas in which the $^{14}$C radiocarbon samples were taken from and the costs associated with dating the samples meant that an age/depth model could not be generated for BAR03 on the lake margin. A second chronology model for the margin of Lake Baraba would have been extremely useful to confirm the age of peat accumulation at the lake margin and also to infer the age of the minerogenic sediments visible below the Holocene peat formation in BAR03 on the lake margin.
7. CONCLUSION

7.1 Summary
This honours thesis was aimed to assess the variability within Lake Baraba, allowing recent trends of significant water loss throughout the TLNP to be understood in the context of long term environmental change within the TLNP. The previous responses of the lake to climatic change throughout the Late-Quaternary, in particular the Holocene record has been inferred through analysis of the sediment record of Lake Baraba. This was achieved via investigating the palaeo-environments of Lake Baraba, examining its sedimentary characteristics (structure and geochemistry) of Lake Baraba and inferring them to previously completed research from Lake Baraba (Black et al., 2006) and Lake Couridjah (Barber, 2018; Gergis, 2000). Sedimentary data included examining the grainsize, age, elemental composition and organic element analysis. The results of the subsequent analysis were compared to other facies found within Lake Baraba and other literature released on palaeo-environments of south-east Australia. Therefore, the following conclusions can be drawn from the research:

1. It is clear due to the variability of sedimentary characteristics (i.e. texture, grainsize, colour and elemental data) that Lake Baraba has undergone significant hydrological change throughout the late-Quaternary, particularly from MIS 3 (~60-24 ka) to present.

2. While the true conditions prior to the Holocene are difficult to infer from the cores BAR02 and BAR03, there is clear evidence suggesting a sustained lacustrine environment in Lake Baraba since the early-Holocene resulting in peat formation and accumulation (~9.3 ka cal. years BP). Although there is some evidence of a variable shore line at the lake margin during the mid-Holocene, this is not seen in the centre of Lake Baraba further indicating the relative stability of the lake water level throughout the Holocene, and cooler drier conditions experienced across south-east Australia didn’t majorly affect the water levels of Lake Baraba.

3. The stratigraphic relationships shown in this study, in combination with research conducted by Barber (2018) indicate that Lake Baraba has undergone multiple different phases from predominantly ‘swampy’ conditions throughout the late-Quaternary, which have in the past connected the two lakes (e.g. pre-Holocene peat layers). Different $^{14}$C dates from the bottom of the peat units in Lake Baraba and Lake Couridjah further indicate separate hydrological and sedimentological process’ occurring, as Holocene peat deposition is inferred to begin in Lake Baraba (~9.3 ka cal. years BP), whilst this date is slightly later in Lake Couridjah (~10.8 ka cal. years BP).

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The TLNP in the context of other peat-forming lakes and swamps in the Sydney Basin reflects a relatively rare system, in which its very unique long-term catchment stability has produced excellently preserved lacustrine sediments and a near constant peat record of peat deposition the last \( \sim 9.3 \) ka cal. years BP. In contrast, many THPSS environments reflect inter-annual hydrological variability which has inhibited the long-term production and preservation of peat (Fryirs et al., 2014), highlighting the unique palaeo-archive of Lake Baraba within the THPSS systems. The lakes are ecologically, scientifically and aesthetically important to the context of the Sydney Basin and south-east Australia despite their apparent long term hydrological variability. Lake Baraba has remained relatively hydrologically stable during the Holocene while the lakes are still subject to changes due to regional climatic changes in the pre-Holocene oxidised sediments of the LGM and interglacial periods followed by stability throughout the Holocene.

**7.2 Future Research**

This student research project has provided some important insights into the palaeo-environmental history of the Thirlmere Lakes, in particular the Holocene peat record of Lake Baraba. Further research is needed however to interpret the pre-Holocene palaeo-environmental history in the centre of Lake Baraba throughout the late-Quaternary. Therefore, in regard to this particular study, further research suggested includes;

1. **Deeper core from the centre of Lake Baraba**

Further stratigraphic interpolation is required in the centre of Lake Baraba, however the core attained from the centre of Lake Baraba (BAR02) was not able to be obtained below the Holocene record. Therefore, a deeper core from the lake centre would allow the stratigraphic relationships seen on the margin of Lake Baraba to be traced through to the middle of Lake Baraba, in turn allowing a better understanding of the previous wet/dry conditions in the centre of Lake Baraba throughout the late-Quaternary.

2. **High clarity organic elemental analysis for the lake centre**

Organic content data from the centre of Lake Baraba would provide a higher resolution differences in the carbon and nitrogen content of Lake Baraba, allowing the palaeo-environments of the driving the different zones of Holocene peat formation to be determined at a higher clarity, and thus a greater precision in the exact peat forming conditions of Lake Baraba can be understood throughout almost the entire Holocene (\( \sim 12-0 \) ka).

3. **Drivers of water level change in the TLNP**
Further research to understand what is causing the significant drop in water level in the TLNP is required to directly determine the anthropogenic impact upon the lake systems. It has been shown that the lakes have been relatively stable throughout for the past 9.3ka cal. years BP, so what has change to cause the fluctuating water level apparent in the modern TLNP.
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APPENDIX A – Rainfall and Lake Level Graphs

The nearest weather station (Buxton approximately 3km away from the TLNP), capturing local rainfall patterns and Ainsley weather station, located 180km away to the South-west, representing regional rainfall patterns (source: Barber, 2018).
APPENDIX B – PCA analysis for sill facies (A) and lacustrine facies (B)

A

B