

DROUGHTS AND FLOODING RAINS:  
A FINE-RESOLUTION RECONSTRUCTION OF  
CLIMATIC VARIABILITY IN WESTERN VICTORIA,  
AUSTRALIA, OVER THE LAST 1500 YEARS.

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Discipline of Geographical and Environmental Studies,  
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2010

## CHAPTER 1 – INTRODUCTION

### 1.1 Introduction

This thesis presents the results of a study examining the short-term (approximately 1500 year) history of climatic change from the south-eastern Australian mainland. In order to achieve this, fossil diatom assemblages from sediment cores retrieved from two crater lakes in western Victoria are examined in fine resolution and interpreted through the application of a diatom-conductivity transfer function developed specifically for use in low salinity lake systems.

### 1.2 Research context

In early November 2006, the then Prime Minister of Australia, John Howard, called an emergency ‘water summit’ with the premiers of the south-eastern states to discuss the impact of the prevailing drought on water availability in the Murray Darling Basin, a key agricultural region for the nation. At this meeting, the attendees were informed that the drought that was afflicting the region was “the worst in 1000 years” (Shanahan and Warren, 2006). This claim proved so startling that it was widely disseminated through the local and national media and was even reported as far away as the United Kingdom (Vidal, 2006). The Prime Minister, however, remained unconvinced of the veracity of the claim because “there are no records [to verify it against]” (Shanahan and Warren, 2006).

This brief exchange between the Nation’s leaders and the press highlights the problems faced by water, agricultural, environmental and social planners in Australia. That is, a lack of knowledge regarding short-term (decadal to sub-decadal) climatic variability beyond the period covered by even the longest instrumental climate records – around 120 years (Nicholls *et al.*, 2006). There are a multitude of palaeoenvironmental and palaeoclimatological studies that have emanated from Australia, though comparatively few provide data at a resolution that can, for example, identify the last occurrence of a drought equivalent to the current event. Knowledge of past climatic variability is important for several reasons, not least of which is determining natural cycles of variability and locating contemporary events within those cycles. Such knowledge can inform on the potential for an increasing or decreasing frequency of climate anomalies and what form those anomalies may take, while also placing the changing Australian

climate of recent decades (Murphy and Timbal, 2008) into a longer-term historical context.

The climate of Australia is highly variable both spatially, because of the range of climate zones on the continent, and temporally, due to the variable and inter-connected influences of major oceanic and atmospheric processes emanating from the Pacific, Indian and Southern oceans and the tropics (Turney *et al.*, 2006). Due to this variability, drought is a common occurrence across much of the continent. The most recent period of drought in the south and east of the country commenced in 1996 and continues to the present (Murphy and Timbal, 2008). Sometimes referred to as ‘The Big Dry’ (Ummenhofer *et al.*, 2009) or ‘The Millennium Drought’ (Bond *et al.*, 2008), it is acknowledged as the most severe drought on record (Murphy and Timbal, 2008), with its unprecedented intensity caused by rising temperatures over recent decades which has increased evaporation rates (Nicholls, 2004; Murphy and Timball, 2008).

The southeast of the mainland is one of the most severely affected regions (Figure 1.1) and, in terms of agricultural production, one of the most important (Murphy and Timball, 2008). This region includes the southern reaches of the countries’ largest river system, the Murray-Darling Basin, which contains 40% of all Australian farms and produces a third of the nations’ food supply (Sandeman, 2008). The majority of this agriculture is heavily dependant on water from the river system and, in 2005-2006, a period when annual average rainfall across the region was in the lowest three decile ranges (AusBOM; climate statement 2006), the agricultural industry accounted for 83% of all water consumed in the basin (Sandeman, 2008). In the study region of western Victoria, inflows into water storage catchments have been significantly reduced, resulting in water storage levels being measured at only 7% capacity in 2006 (Murphy and Timbal, 2008). The drought has also impacted on wetland resources across western Victoria with numerous lakes drying out. Among these are several Ramsar-listed wetlands, some of which, in 2002, were classified as being ‘permanent saline’ (Lake Bookar, Lake Gnarpurt) or ‘permanent freshwater’ (Lake Terangpom) wetlands (Victorian Government, 2002).

The primary cause of the current drought, and other iconic droughts of the 20th century, has been identified as a combination of the influences of *El Niño* and the positive mode of the Indian Ocean Dipole (IOD) (Ummenhofer *et al.*, 2009). The converse also

appears to be the case with the co-occurrence of *La Niña* and the negative mode of the IOD identified as a driver of positive rainfall anomaly in the southeast of the country (Meyers *et al.*, 2007).

The relatively recent discovery of the IOD (Saji *et al.*, 1999), and the recognition of its contribution to rainfall anomalies in various regions in Australia (e.g., Smith *et al.*, 2000; Ashok *et al.*, 2003; Meyers *et al.*, 2007; Ummenhofer *et al.*, 2009), demonstrates the growing body of knowledge about sub-decadal scale climate variability in Australia. However, knowledge of similar scale variability of past climates is, by and large, limited to the period of instrumental records or restricted geographic regions such as western Tasmania (Cook *et al.*, 1991, 1992, 2000). In order to gain an understanding of climate in its entirety, there is a need to increase the focus on documenting short-term, high-amplitude, climatic variability beyond what is known from instrumental records.

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This figure is included on page 3 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 1.1: Observed national mean rainfall anomalies for the period December 2003 – November 2006. Source; Australian Bureau of Meteorology website.*

*Redrawn from Hunt (2009)*

Two decades ago, Wasson (1990) concluded that “[in terms of palaeoclimate studies], the last 2000 years has been largely ignored in the Australian context.” In the years since, relatively little has changed. With the notable exception of the Tasmanian

temperature reconstructions by Cook *et al.* (1991; 1992; 2000), the studies that do exist are generally either of a low temporal resolution (e.g., Holdaway *et al.*, 2002; Pollack *et al.*, 2006 and, to a lesser extent, Mooney, 1997) or cover only a brief window of this time period (e.g., Hendy *et al.*, 2002a). In fact, at present, there are no sub-decadal resolution, millennia-length studies from the Australian mainland.

The lack of finely-resolved data from the Southern Hemisphere has been noted in reviews of climate studies (e.g., Mayewski *et al.*, 2004; Ljungqvist, 2009), hemispheric temperature reconstructions (e.g., Mann and Jones, 2003; Mann *et al.*, 2008) and, recently, by the Intergovernmental Panel on Climate Change (Jansen *et al.*, 2007). Partly as a consequence of this ‘research gap’, there exists a large discrepancy in the knowledge and understanding of climatic variability of the last two millennia in the Northern Hemisphere and that which exists in the Southern Hemisphere and, more specifically, in Australia. In particular, the two most notable periods of climatic anomalies during the last 1500 years – the so called ‘Little Ice Age’ (LIA – *ca.* 1350-1850 AD) and the ‘Medieval Warm Period’ (MWP – *ca.* 800-1200 AD) – are widely documented in Northern Hemisphere studies, though the impacts of these periods, if any, and the timing of their onset and duration, have not been identified in mainland Australian studies. Generally characterised as being a period of elevated temperatures (Lamb, 1965), the MWP occurred when global boundary conditions (e.g., insolation, ice-cover, sea-levels) were similar to present and may provide, therefore, a potential analogue (Bryson, 1985) to future climates under the influence of global warming. Hence, knowledge of climatic variability during such times may provide a possible insight into future climate regimes.

### **1.3 Aims of the study**

In recognition of the paucity of, and need for, highly-resolved palaeoclimate studies from Australia, the aim of this project is to reconstruct, at sub-decadal resolution, climatic change from south-eastern mainland Australia over the last millennium or more. In order to achieve this, sediments from two volcanic crater lakes in western Victoria were examined at high-resolution for changes in the species composition of diatoms. The importance of these changes, and their relationship to climate, are determined through the application of a diatom-conductivity transfer function. More specific aims of the study are:

1. To identify high frequency climate variability and determine if and how this has changed over time.
2. To identify low frequency climate variability in order to determine the presence or absence of long-term cycles and, if these cycles are evident:
  - a. contextualise contemporary climates within these and,
3. To identify the frequency, intensity and duration of previous drought episodes in order to:
  - a. Compare the current drought to previous episodes
  - b. Provide evidence of long-term regional drought history
4. To compare and contrast results with other relevant studies in order to provide a regional synthesis and determine possible driving mechanisms behind identified changes.
5. To characterise the climate in western Victoria during the periods commonly associated with the LIA and MWP and assess this in terms of the debate around the spatial extent of climate phenomena observed during these periods.

The research plan has been specifically structured towards achieving these aims. Foremost within this is the identification of suitable study sites that are climatically sensitive, contain a continuous record of sediment deposition, have diatoms preserved in the sediment and, crucially, have a sedimentation rate that is amenable to achieving the aim of sub-decadal sampling resolution. As elaborated in the body of the thesis, an existing diatom-salinity transfer function for the region (Gell, 1997) was skewed to salinities considerably higher than the study lakes. Therefore, a new diatom-conductivity transfer function was developed, tailored specifically for application to the study lakes. The aims of this aspect of the project are:

1. To develop a statistically robust diatom transfer function that can quantitatively reconstruct conductivity from low-conductivity lakes,
2. To evaluate transfer function performance against similar models,
3. To validate the transfer function by comparing inferred-conductivity changes over the historical period to instrumental climate records,
4. To apply the model to fossil diatom assemblages from the study sites in order to reconstruct conductivity change.

#### **1.4 Thesis outline**

In the remainder of the thesis, a detailed contextual discussion is provided in **Chapter 2**. Here, millennial-scale climate variability since the mid-Holocene is discussed in terms of studies from both the Northern and Southern Hemisphere, followed by a review of centennial-scale variability of the last two millennia, with a specific focus on the Australasian region. This is followed by an assessment of the utility of diatoms for palaeolimnological studies and, more specifically, fine-resolution palaeoclimate reconstructions. **Chapter 3** provides a background to the study region of the western plains of Victoria, and the study sites, Lake Elingamite and Lake Surprise. In **Chapter 4**, field, laboratory and statistical methods are presented, while the results of laboratory and statistical analyses are presented in **Chapter 5**. In **Chapter 6**, the performance of the transfer function is evaluated by comparing salinity reconstructions against the instrumental climate record. Conductivity reconstructions from both lakes are compared to determine differing lake responses, while comparisons with other relevant studies determines possible driving mechanisms behind the identified changes. **Chapter 7** concludes the thesis by reviewing, and assessing, the various components of the project, summarising the research findings, identifying any implications for further research and determining if the stated project aims have been achieved. It will be demonstrated that the findings of this project have significantly contributed to, and advanced, the current body of knowledge about past climate variability in south-eastern Australia.

## CHAPTER 2 – CLIMATIC VARIABILITY, HIGH-RESOLUTION PROXY ANALYSIS AND PALAEO LIMNOLOGY

### 2.1 Introduction

Palaeoecological studies provide evidence of past environments of a location or region over extended timescales, thereby extending and testing knowledge of climatic variability and the nature of environmental response. Comparisons between fossil evidence and modern analogues provide the basis for reconstructing these environmental characteristics and can give an indication of past climates whilst also providing the opportunity to view current climatic conditions within an historical context. ‘Stand-alone’ studies, from a single site using a single proxy, can offer significant insight into past environments (e.g., Kershaw, 1974), though correlation and comparison between studies from the same region, spanning the same period, can add both support for interpretations and a more precise spatial perspective to the results (e.g., Kershaw *et al.*, 1993; Moss and Kershaw, 2000).

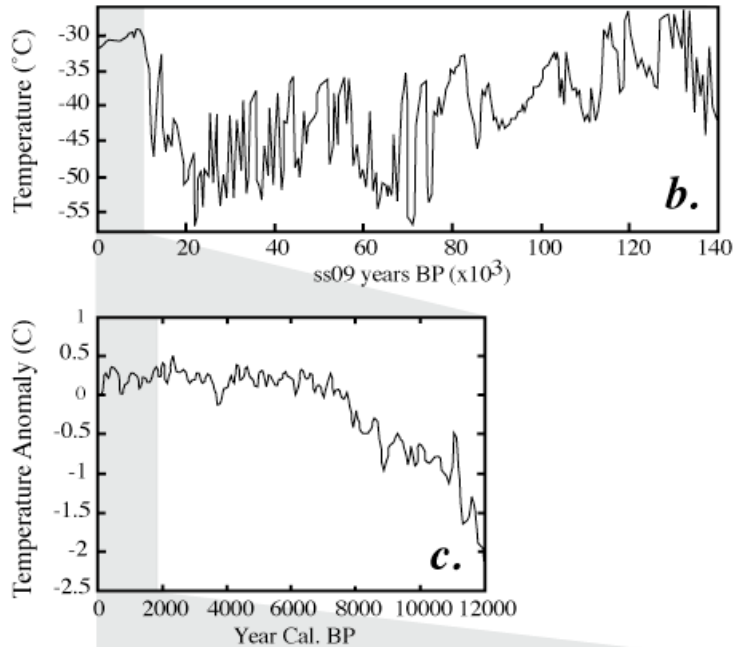
The integration of further lines of evidence, a multi-proxy approach, can produce greater detail and more evidence of the impact of identified perturbations. Furthermore, where the records of some proxies are discontinuous or sampled at a lower resolution, other proxies may remain and ‘fill the gaps’ (e.g., Mooney, 1997). Conversely, the concurrent presence of multiple proxies can provide increased confidence in the interpretation (e.g., Bertrand *et al.*, 2008). Ultimately, where the combination of proxy evidence provides a robust interpretation, the results can become a reference against which further studies from the same region, or further afield, can be compared. Lake Keilambete, in western Victoria, is an example of such a site and illustrates the benefits that can accrue with the addition of further lines of investigation. Here, analysis of sediment stratigraphy (Bowler and Hamada, 1971), pollen (Dodson, 1974), sedimentary geochemistry (Bowler, 1981) and ostracod shell chemistry (Chivas *et al.*, 1986) have combined to provide a Holocene lake-level and climate history of the region. This, in turn, has informed numerous further climate studies of the region (e.g., Edney *et al.*, 1990; Gell *et al.*, 1994; Dodson, 2001; Tibby, 2003; Kershaw *et al.*, 2004b) and beyond (e.g., Markgraf *et al.*, 1986; Holdaway *et al.*, 2002; Stanley and De Deckker, 2002; Tibby and Haberle, 2007).



The comparison of records from different regions enables the examination of the disparity and/or synchronicity that exists between identified climatic events. This, in turn, facilitates assessment of the relative impact of regional or global climate drivers on a given study region and informs understanding of the nature and impacts of climatic change and variability (e.g., Mann *et al.*, 2003, 2008; Wanner *et al.*, 2008). Natural climate change and variability is driven by numerous overlapping climatic cycles which operate on temporal scales ranging from seasonal to multi-millennial (Figure 2.1). A strength of palaeoenvironmental research lies in the ability to identify these cycles and determine the manner in which varying environments respond to, and recover from, such perturbations (e.g., Dodson and Ono, 1997; Mayewski *et al.*, 2004; Bertrand *et al.*, 2008).

Temporal scale is therefore a key issue, though the degree to which this can be investigated is often determined by the study site and/or the proxies employed. A greater temporal scale usually comes at the expense of sample resolution. Marine sediment cores for example, can provide very long and relatively detailed sequences over time periods of  $10^5$  or  $10^6$  years (e.g., Shackleton and Opdyke, 1977; Lisiecki and Raymo, 2005), though due to the generally low sedimentation rate and the effects of bioturbation, these cores are normally unable to provide insights into fluctuations at a decadal or centennial scale. Conversely, dendrochronological studies can provide very high resolution records of annual or seasonal variation (e.g., Cook *et al.*, 1991, 1992, 2000; Briffa, 2000; D'Arrigo *et al.*, 2008), though continuous studies can only be extended over a period of hundreds to thousands of years.

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*Figure 2.1: Climatic change and variability at timescales ranging from 1Mya to 2000 years. Note the changing values on vertical axes. **a.** Lisiecki and Raymo (2005) a stack of 57 correlated, globally distributed, benthic  $\delta^{18}\text{O}$  records showing the procession of glacial/interglacial stages over the last million years; **b.** Johnsen et al. (2001) Greenland temperature variations over the last 140ka from GRIP  $\delta^{18}\text{O}$  records and borehole thermometry; **c.** Davis et al. (2003) European annual temperature anomalies of the Holocene reconstructed from pollen data; **d.** Mann and Jones (2003) smoothed global temperature reconstruction (blue) and instrumental records (red) since 200 AD, showing temperature anomalies away from the 1961-1990 instrumental reference period mean (dashed line). The general timing of the periods which form the basis for thesis aim number 5, the Little Ice Age (LIA) and Medieval Warm Period (WMP), are highlighted.*

Hence, a combination of long-term, low-resolution studies and short-term, high-resolution studies is needed to gain an overall illustration of climatic change and variability at differing timescales. However, determining the timing of the onset of climatic changes, assessing the degree of synchronicity between regions and/or hemispheres and evaluating the relative effects of the mechanisms forcing climatic variability, requires such studies to have a relatively equal spatial distribution. At present, this is not the case, with far fewer studies emanating from the Southern Hemisphere than the Northern Hemisphere (Jansen *et al.*, 2007), making inter-hemispheric comparisons difficult (Mann and Jones, 2003; Mann *et al.*, 2008; Wanner *et al.*, 2008). This is perhaps a contributing factor to debate about whether prominent millennial-scale (Bond *et al.*, 1997, 1999, 2001; Broecker and Hemming, 2001; Crowley, 2002) and centennial-scale (Bradley *et al.*, 2001; Broecker, 2001) climatic cycles evident in Northern Hemisphere records exist in the Southern Hemisphere and, more specifically, the extent of their impact and their degree of synchronicity with Northern Hemisphere signals.

## **2.2 Millennial-scale climate change**

The dominant climate forcing mechanisms at a millennial scale are related to astronomical variations in the eccentricity, obliquity and precession of the Earth's orbit. Collectively, these are referred to as Milankovitch cycles after the Serbian mathematician who first identified them, Milutin Milankovitch (Imbrie, 1982; Berger, 1988; Hartmann, 1994). Each of the cycles operates on a different time span ranging from approximately 19,000 years (precession) to approximately 100,000 years (eccentricity). While this theory was proposed by Milankovitch, it was confirmed by Hays *et al.* (1976:1131) who concluded that "changes in the Earth's orbital geometry are the fundamental cause of the succession of Quaternary ice ages" (see Figure 2.1a).

Within the last glacial period (Figure 2.1b), at a 'sub-Milankovitch' scale, are further examples of cyclic climatic variability. First identified in the Camp Century Greenland ice core by Dansgaard *et al.* (1984), and then confirmed by Johnsen *et al.* (1992), Dansgaard *et al.* (1993) and Grootes *et al.* (1993), the last glacial period contains between 21 (Labeyrie *et al.*, 2003) and 24 (Dansgaard *et al.*, 1993) rapid, large amplitude temperature oscillations commonly referred to as Dansgaard-Oeschger (D-O) events. The D-O events take the form of rapid warming of several degrees over the

space of several decades, followed by a gradual cooling over a period of several centuries to millennia (Alley, 2000). These events occur approximately every 1470 years (Schultz, 2002; Rahmstorf, 2003), though whether this re-occurrence is cyclic is still under debate (Schulz, 2002; Schulz *et al.*, 2004; Braun *et al.*, 2005; Ditlevsen *et al.*, 2007). The D-O events have been linked with the discharge of massive ice rafts from the Laurentide ice sheet (Heinrich, 1988; Timmermann *et al.*, 2003), the so-called ‘Heinrich’ events, and smaller ice rafting events from other ice sheets (Bond and Lotti, 1995). While most of the evidence of D-O events emanates from the North Atlantic region (Voelker *et al.*, 2002), there is evidence of their impact from lower latitudes in the Northern Hemisphere (Hendy and Kennett, 1999; Sachs and Lehman, 1999; Peterson *et al.*, 2000; Hendy *et al.*, 2002b) and various locations in the Southern Hemisphere (Charles *et al.*, 1996; Kanfoush *et al.*, 2000; Hinnov *et al.*, 2002). Notably, the signature of D-O events in the Antarctic ice-core record is very muted with oscillations showing, at best, only 10% the amplitude of records from the north (Hinnov *et al.*, 2002).

The occurrence of these events during the last glacial period places them beyond the scope of this project, however there are suggestions that the D-O events do continue, though in muted form, through the Holocene period in the form of the so-called ‘Bond Events’ (Bond *et al.*, 1997; 1999; 2001). These events take the form of cool periods and were identified through the fine-resolution analysis of ice-rafted debris and  $\delta^{18}\text{O}$  of foraminiferal remains in deep ocean cores in the North Atlantic (Bond *et al.*, 1997).

There is still debate as to whether Bond events are the interglacial equivalent of the D-O cycles (Shuman *et al.*, 2005; Avery and Singer, 2006; Buetikofer, 2007). Even after an extensive review of mid- to late-Holocene climatic change, Wanner *et al.* (2008:1819) could only conclude that Bond events “may or may not be analogous to Dansgaard-Oeschger events”. There is also debate as to the claim by Bond *et al.* (1997, 1999, 2001) that the occurrence of the events are cyclic in nature. This assertion is not supported by the analyses of Buetikofer (2007), or the climate models of Braun *et al.* (2005), while Wanner *et al.* (2008:1819) conclude that “there is scant evidence ... for the cyclicity of climate variations on this timescale”. However, regular ~1500 year cycles have been found in North America (Viau *et al.*, 2002), western Canada (Yu *et al.*, 2003), the Arabian sea (Berger and von Rad, 2002), central Africa (Stager *et al.*, 2003), Mexico (Castiglia and Fawcett, 2006), monsoonal Asia (Gupta *et al.*, 2003; Hong *et al.*, 2003)

and the east coast of Australia (Skilbeck *et al.*, 2005). Indeed, Singer and Avery (2005:17) argue that “the geographic range and variety of evidence for a 1,500-year climate cycle is too great to dismiss”.

Irrespective of the debates surrounding cyclicity, the actual presence of Bond events throughout the Holocene is undisputed. There have been nine such events identified (Bond *et al.*, 1997; 2001) with the Little Ice Age (LIA), ‘Bond event zero’, being the most recent (Figure 2.2). Apart from the LIA, these include other well documented periods of significant climatic change in the North Atlantic region such as the ‘8.2 ka event’ (Alley *et al.*, 1997) and the ‘2.8 ka event’ (Swindles *et al.*, 2007).

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*Figure 2.2: Numbered Bond events identified from stacked records of North Atlantic ice rafted debris. Redrawn from Bond et al. (2001)*

In terms of millennial-scale climate change, the general trend evident in the stacked records of Bond *et al.* (2001) is one of a gradually warming climate (Figure 2.2), though this is at odds with several other Northern Hemisphere records, especially when focussing on the climate of the last 6000 years<sup>1</sup>. Indeed, a number of proxy-reconstructed temperature records of this time period from the North Atlantic region show an almost linear decrease in temperature, approximately until the end of the Little Ice Age (*ca.* AD 1850) (Fisher *et al.*, 1994; Seppä and Birks, 2001; 2002; Vinther *et al.*,

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<sup>1</sup> The time-slice of 6000 years will be the focus of this discussion as this is a period lacking the climatic influence of large continental ice sheets and mass freshwater discharges from them, while is also a period of relatively stable sea levels (Edwards, 2006; Wanner *et al.*, 2008). Climatically, these factors ensure a relatively stable period making inter-hemispheric and inter-regional comparisons more straightforward.

2006). There is also evidence of gradually advancing glaciers across the European Alps over this time (Wanner *et al.*, 2008). This trend of general cooling is supported by sea-surface temperature reconstructions from the north of Iceland (Andrews *et al.*, 2001; Bendle and Rosell-Melé, 2007), the Norwegian Sea (Koç *et al.*, 1993; Calvo *et al.*, 2002) and the North Atlantic Ocean (Marchal *et al.*, 2002; Rimbu *et al.*, 2003; Andersen *et al.*, 2004), which show a similar trend.

Northern Hemisphere records from further afield are also in general agreement with these observations. Viau *et al.* (2006) reconstructed mean, continental, summer temperature for North America which clearly demonstrates a trend of declining summer temperatures in the late Holocene. Davis *et al.* (2003) undertook a similar pollen-based study for northern, central and southern Europe and also demonstrated declining summer temperatures since the mid-Holocene for northern Europe, though central Europe shows relatively stable temperatures while the southern Europe region shows increasing summer temperatures for this period. Hydrologically, the Mediterranean region became progressively more arid after the mid-Holocene (Harrison and Digerfeldt, 1993; Bar-Matthews *et al.*, 2003; Davis *et al.*, 2003), as did regions of China and Tibet (Feng *et al.*, 2006; Yu *et al.*, 2006) and North America (Valero-Garcés *et al.*, 1997; Thompson and Anderson, 2000).

It is widely noted that decreasing summer insolation at the high latitudes of the Northern Hemisphere (Figure 2.3), through the late-Holocene, was the primary driving mechanism behind this pervasive mid- to late-Holocene cooling trend (Berger and Loutre, 1991; Wright *et al.*, 1993; Koç and Jansen, 1994; Marchal *et al.*, 2002; Wanner *et al.*, 2008). Changes in insolation in the tropics and sub-tropics display a far more muted trend (Wanner *et al.*, 2008), though there is widespread evidence of climatic change which takes the form of an almost uniform trend of increasing aridity through the late Holocene. This is perhaps most evident in tropical Africa where there was increasing rainfall variability and aridification (deMenocal *et al.*, 2000a; Gasse, 2000, 2001; Nguetsop *et al.*, 2004), but similar conditions have been identified in Central America (Hoddell *et al.*, 1995; Haug *et al.*, 2001) and Asia (Wanner *et al.*, 2008). This trend has been attributed to the gradual southward migration of the Inter Tropical Convergence Zone (ITCZ) since approximately 5400 years ago as evidenced in fluctuations in the deposition of terrigenous materials offshore from Venezuela (Haug *et al.*, 2001).

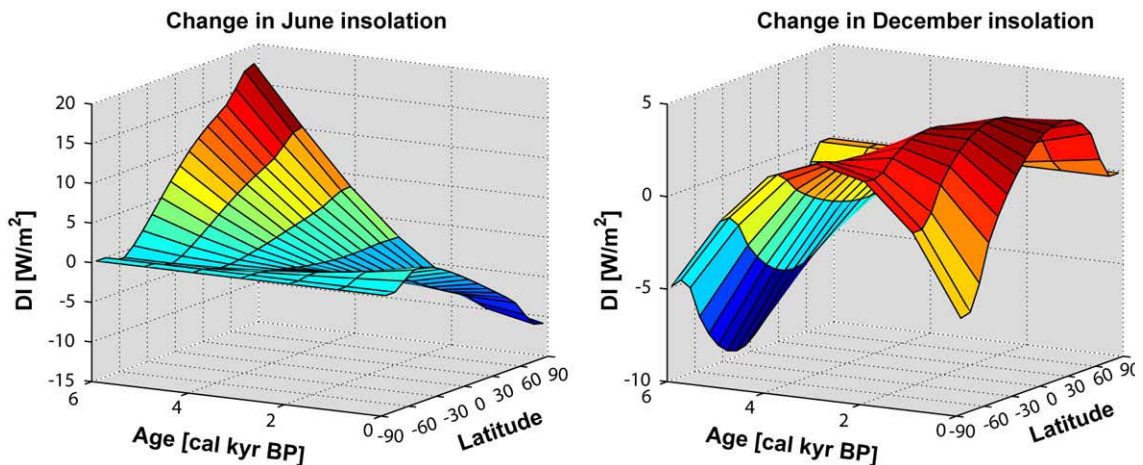


Figure 2.3: Calculated deviations of insolation from long-term mean ( $W/m^2$ ) over the last 6000 years for northern and southern summers. Note differing scales on the vertical axes. From Wanner *et al.* (2008)

Northern tropical and sub-tropical sea-surface temperatures (SSTs) do not demonstrate such a coherent signal. Kim *et al.* (2004) compared numerous SST records of the last 7000 years from the North Atlantic, North Pacific and Indian Oceans and found a general cooling trend evident in sub-tropical Atlantic records and a general warming trend in the sub-tropical Pacific and Indian Ocean records. The cooling trend of the sub-tropical Atlantic has been attributed to a possible southward movement of cooler temperate, or sub-polar, waters or increased regional upwelling (deMenocal *et al.*, 2000b) and displays a close connection to North Atlantic climate records (deMenocal *et al.*, 2000b; Kim *et al.*, 2004). The Indian Ocean SST record may be driven by changes in monsoon variation (Jung *et al.*, 2004), which is again linked to ITCZ migration (Haug *et al.*, 2001), whereas the “seesaw” effect between Atlantic and Pacific SSTs may be driven by a complex interaction involving atmospheric circulation and sea level pressure between the North Atlantic Oscillation and the North Pacific equivalent, the Pacific North American circulation pattern (Kim *et al.*, 2004).

It is clear that south of the equator the records from the Pacific Ocean indicate a different climate. In contrast to the general warming trend outlined by Kim *et al.* (2004), Stott *et al.* (2004) found a cooling trend in SSTs from the equatorial western Pacific since the mid-Holocene which they derived from foraminiferal Mg/Ca ratios. This is

supported by the findings of Gagan *et al.* (1998) who identified a 1°C cooling since ~5350 years ago from Sr/Ca and <sup>18</sup>O/<sup>16</sup>O ratios in Great Barrier Reef coral records, indicating a possible prevalence of *El Niño* conditions. Over the same time frame Abram *et al.* (2007) demonstrate a long cooling period in the Indian Ocean coupled with a weakening Asian monsoon and strengthening *El Niño* activity.

A proposed increase in ENSO activity since the mid-Holocene is supported by the findings of several studies including that of Moy *et al.* (2002) from lake sediments in southern Ecuador. Here, an increase in the frequency of rainfall events since the mid-Holocene, though especially within the last 2000 years, is attributed to anomalously warm SSTs off the coast of Ecuador initiating convection. While this is countered by pollen studies from El Salvador that indicate a gradual drying trend through the late-Holocene (Dull, 2004), the results of this study may be due to the impact of local influences as numerous studies also identify a pattern of increasing ENSO variability since the mid-Holocene, with a tendency towards a prominent *El Niño* pattern (Shulmeister and Lees, 1995; Rodbell *et al.*, 1999; Shulmeister, 1999; Andrus *et al.*, 2002; Riedinger *et al.*, 2002; Donders *et al.*, 2007), possibly driven by orbital forcing (Clement *et al.*, 2000). To a large degree, this explains the trend of increasing effective precipitation that is present in records from sub-tropical and tropical South America (Grojean *et al.*, 1997; Thompson *et al.*, 1998; Baker *et al.*, 2001; Grojean *et al.*, 2003; Tapia *et al.*, 2003) which is in direct contrast to the widespread trend of increasing aridity in the northern tropics and sub-tropics described above.

As opposed to increasing effective precipitation in the south-eastern Pacific, a predominance of *El Niño* conditions should be reflected in increasing aridity in the south-western Pacific. Evidence of such is provided from several sites in Indonesia and Papua New Guinea (Haberle *et al.*, 2001) where an increase in biomass burning is found in the fossil charcoal and pollen records that cannot be solely attributed to human influence. Similarly, in northern Australia, where the ENSO phenomenon has a major effect on rainfall variability (Nicholls, 1988; Nicholls and Kariko, 1993), Tibby and Haberle (2007) found evidence of falling lake levels and increased climatic variability in the northern tropics, particularly after 3300 cal. yr BP. This reflects the findings of Kershaw and Nix (1988), who identified a decreasing mean annual precipitation commencing around the same time, and Rowe (2007), who identified a vegetation shift away from rainforest taxa towards open sclerophyll forest in the tropical Cape York



area of Australia at around 3000 radiocarbon years BP. Also, from the tropical north of the country, Shulmeister and Lees (1995) identified the onset of a highly variable ENSO-dominated climate at around 4000 years BP from pollen records.

The impacts of increased ENSO variability and an *El Niño* dominated climate are also evident from lower latitudes. In southeast Australia, Bowler (1981) finds evidence of enhanced climatic variability after 3750 cal. yr BP, as do Tibby *et al.* (2006) from the same region. Similarly, from a compilation of southern Australia Holocene charcoal records, Kershaw *et al.* (2001) identify an increase in landscape burning occurred after 4000 years BP. This may, in part, be due to an increasing Aboriginal population after 5000 cal. yr BP (Smith *et al.*, 2008), though several other studies from the region also indicate a pattern of decreasing effective precipitation since 5000 years BP (e.g., Singh, 1981; D'Costa *et al.*, 1989; Singh and Luly, 1991; McKenzie and Kershaw, 1997; Kershaw, 1998). Harrison and Dodson (1993) argue that this trend may be due to an expansion of the sub-tropical high-pressure belt limiting moisture bearing westerly air flow in the winter months. However, several others (Singh and Luly, 1991; Williams, 1994; Nanson *et al.* 1995; Shulmeister, 1999) suggest that a northwards contraction of the Australian monsoon, which would have the effect of reducing the occurrence of summer monsoonal incursions to the south of the continent, might be the underlying cause. Increasing summer insolation (Figure 2.3) may also be a contributing factor by increasing evapotranspiration (Shulmeister, 1999).

Further south in Tasmania, Markgraf *et al.* (1986) and Anker *et al.* (2001) suggest that the late Holocene was generally drier and cooler than the mid and early Holocene. Similarly, in New Zealand, a shift toward more drought and frost tolerant vegetation occurred after 5000 BP (Shulmeister, 1999) in line with glacial expansion (Gellatly *et al.* 1988), though the prevalence of storms may have been greater (Gomez *et al.*, 2004). The gradual drying and cooling trend has been attributed to the increase in westerly circulation since the mid-Holocene which, in turn, is linked to changes in seasonality and the precessional cycle (Shulmeister, 1999). Markgraf (1993) proposes that this strengthened westerly circulation is the mechanism behind increasing rainfall in southern South America over the late-Holocene, evidence of which was also found in records from both Patagonia (Pendall *et al.*, 2001) and Chile (Sterken *et al.*, 2008). The increased moisture availability also led to glacier advances in the major ice fields of

Patagonia after 6000  $^{14}\text{C}$  yr BP, with the exception of an arid phase between 3600 and 3000  $^{14}\text{C}$  yr BP (Glasser *et al.*, 2004).

In the southern high latitudes, comparisons between sites in Antarctic show varying results. Masson *et al.* (2000) provide a summary of millennial-scale climate variability derived from isotopic changes in eleven ice cores from contrasting Antarctic locations. The results demonstrate a complex interaction between temperature, elevation, continentality and oceanic influence though, in general, a general cooling trend is evident in all but one site over the last 5000 years.

### 2.2.1 Summary

In summary, millennial scale climatic change over the last 6000 years has differed significantly in differing regions. Decreasing summer insolation in the Northern Hemisphere, driven by orbital forcing, has led to a measured cooling trend across the mid-to-high latitudes and gradual glacier advances across Europe. In the northern tropics and sub-tropics, a gradual southward migration of the ITCZ led to decreasing rainfall across areas of Africa and central America and variation in the monsoon system (Wanner *et al.*, 2008). In the Southern Hemisphere, there has been a marked increase in ENSO variability since 5000 BP and a tendency towards *El Niño* conditions of a warm eastern Pacific and a cooler western Pacific (e.g., Moy *et al.*, 2002). This has led to a gradual drying trend in parts of Indonesia, Papua New Guinea, Australia and New Zealand, especially over the last 4000 years. In combination with the warm eastern Pacific, the increasing westerly winds, also driven by orbital forcing, have delivered more moisture to southern South America leading to glacial advances across the ice fields of Patagonia (Glasser *et al.*, 2004). In the southern high latitudes, a general cooling trend is evident over the last 5000 years (Masson *et al.*, 2000).

Therefore, apart from a cooling trend evident in the high latitudes of both hemispheres, there are no obvious coherent global, or even hemispheric-scale, trends apparent since the mid Holocene (Wanner *et al.*, 2008). At a regional scale there are areas of increasing aridity and areas of increasing humidity, though even within individual regions there are studies that contradict the general pattern. Yet, superimposed upon the millennial trends, there are numerous shorter, decadal- to centennial-scale changes evident. The

degree to which these fluctuations have a global, hemispheric or regional signature are addressed in the following section.

### **2.3 Centennial-scale climatic variability**

As noted above, the climate of the mid- to late- Holocene has been characterised by regular, possibly cyclic, fluctuations which are widely referred to as ‘Bond Events’ (Bond *et al.*, 1997; 1999; 2001) (Figure 2.2). These, and the intervening warmer periods, generally correspond with significant periods of environmental and, at times, cultural change (Weiss *et al.*, 1993; Hodell *et al.*, 1995; Cullen *et al.*, 2000; deMenocal, 2001; Fagan, 2002). Indeed, several of these episodes – the Iron Age Cold Epoch, the Roman Warm Period, the Dark Ages Cool Period and the Medieval Warm Period – have been assigned popular names based on the European cultural period with which they loosely correspond.

Yet the pattern of the Bond Events is not necessarily expressed across the globe. Mayewski *et al.* (2004) studied approximately 50 globally distributed Holocene palaeoclimate records from which they identify six periods of rapid climate change that had global signatures, only three of which (6000-5000, 3500-2500 and 600-150 cal.yr.BP) correspond to identified North Atlantic Bond Events. Of the six periods identified by Mayewski *et al.* (2004), the two that are most pertinent to this project are those that fall within the last two millennia. Identified by Mayewski *et al.* (2004) as occurring at 1200-1000 and 600-150 cal yr BP, these periods are commonly referred to as the Medieval Warm Period (MWP) and the Little Ice Age (LIA), respectively. These are the two most prominent periods of climatic change over the last two millennia and form the focus of the following discussion.

#### 2.3.1 Climate of the last two millennia

Over recent decades the climate of last two millennia has been the subject of intense research (Jones *et al.*, 2009), though, at first, the focus on 2000 years as a time frame of study may appear purely arbitrary. Ideally, the longer the study period and the finer the sampling resolution, the greater the likelihood of identifying trends, cycles and variability within the data. However, as Lotter and Sturm (1994:311) state, “... such high-resolution paleo-studies are very labour-intensive, [therefore] they are usually only

carried out for specific time-windows containing significant changes in ... parameters of interest.” The last 2000 years contain two such ‘changes’ in the form of climatic events of differing impacts; the so-called Medieval Warm Period, generally the warmest episode of this period (pre- ~1900), and the Little Ice Age, generally the coolest episode (Mann *et al.*, 2009). The presence of two contrasting climate regimes provides the potential to investigate the environmental impacts of these regimes and, given their temporal proximity to the present day and the availability of high-resolution proxies, to do so in great detail. In addition, these events occurred at times when global boundary conditions (e.g., insolation, global ice cover, sea-levels) were relatively similar to the contemporary period and, therefore, provide more appropriate points of comparison than other earlier periods. In terms of understanding the current warming trend then, the MWP in particular may provide at least a partial analogue (Bryson, 1985).

In 1965, H.H. Lamb first articulated the possibility that there had been a prolonged period of warmth during ‘medieval’ times (Lamb, 1965). Lamb compared results from numerous studies and formulated indices of ‘winter mildness or severity’ and ‘summer wetness’. He concluded that, for a period from approximately AD 1100-1200, there was a period of warmer, drier summers and milder winters, with long term temperature averages in England being around 1.2 - 1.4°C above ‘average’. Lamb does not elucidate what constitutes ‘average’, though it could possibly have been the 1931-1960 mean (Bradley *et al.*, 2003a). Lamb labelled this warm period as the ‘Medieval Warm Epoch’ and, while he alluded to the possibility of this being a global phenomenon (citing studies from as far afield as Antarctica, Patagonia and New Zealand), he did not extrapolate his estimates any further than western Europe and, more specifically, England.

While the findings of Lamb (1965) were limited in their geographical context, they have spawned numerous studies and the terms ‘Medieval Warm Epoch’ (e.g., Lamb, 1965; 1969), ‘Medieval Warm Period’ (e.g., Hughes and Diaz, 1994; Keigwin, 1996; Pfister *et al.*, 1998; Piva *et al.*, 2008) or ‘Little Climatic Optimum’ (e.g., Varekamp *et al.* 1992; Wang, 1992; Petersen, 1994) have been widely used in studies from all over the world. While there is now a large body of research on the topic, there remains little agreement on either the timing or the climatic impact of the MWP on hemispheric or even regional scales. A review of the evidence suggesting a prolonged warm phase in Europe, for example, concluded that southern Europe did not experience any significant winter

warmth until the mid-fourteenth century, while western Europe experienced warm springs and dry summers throughout most of the thirteenth century (Alexandre, 1987; cited in Hughes and Diaz, 1994). At centennial time-scales, reconstructed surface temperatures of the MWP (defined as 950 – 1250 CE) suggest a cooling over parts of northern and eastern Europe and a warming around the Mediterranean (Mann *et al.*, 2009).

Still, the concept of a Medieval Warm Period has been widely adopted and there remains a general belief that the global climate of the last 1000 to 1500 years included a warm period of several centuries duration (e.g., Soon and Baliunas, 2003). This assumption is countered by Hughes and Diaz (1994) who question if the Medieval Warm Period truly was a “globally coherent pattern of century scale variation”. In their analysis of available studies they conceded that there was (at the time of publication) relatively sparse worldwide data around the turn of the first millennium and stated that ...

*“it is impossible at present to conclude from the evidence gathered ... that there is anything more significant than the fact that in some areas of the globe, for some part of the year, relatively warm conditions may have prevailed ... [which] does not constitute compelling evidence for a global ‘Medieval Warm Period’”* (Hughes and Diaz, 1994:136).

Despite a plethora of studies since then, the debate continues. The results of Mann and Jones (2003) for instance, do not indicate any significant warm period around the time in question from either the northern or Southern Hemisphere (though they do acknowledge a paucity of Southern Hemisphere data limits the accuracy of their reconstruction). On the other hand, Goosse *et al.* (2005) found that the MWP was “a hemispheric-scale phenomenon, at least”, while Soon and Baliunas (2003) claim that the MWP was a climatic anomaly which had a “worldwide imprint”.

It is certainly clear that numerous studies examining the climate of the last two millennia do find some evidence of a climatic anomaly of some type which falls within the broad European ‘medieval period’ of AD 500 – 1500, though this is not always a temperature anomaly. Therefore, in terms of descriptive terminology, the labels ‘Medieval Warm Period’, ‘Medieval Warm Epoch’ and ‘Little Climatic Optimum’ can be misleading due to the emphasis on temperature as the main defining characteristic. Numerous studies have identified greater hydrological anomalies than temperature anomalies at this time (e.g., Leavitt, 1994; Laird *et al.*, 2003; Lachniet *et al.*, 2004;

Daniels and Knox, 2005; Kondrashov *et al.*, 2005; Ngomanda *et al.*, 2007) and, in this regard, Stine (1994) argues that ‘Medieval Climatic Anomaly’ is a more appropriate term. While this term is becoming more widely used (e.g., Raab and Larson, 1997; Li *et al.*, 2000; Hu *et al.*, 2001; Mann *et al.*, 2009) it is yet to be universally adopted.

It is worth noting though, that the removal of temperature as the defining characteristic of the period also has wider implications. The apparent ‘warmth’ of the MWP has been ‘grist to the mill’ for opponents of action on global warming (Bradley *et al.*, 2003b). The line of reasoning proposed is that if, in the absence of elevated greenhouse gasses, the MWP was as warm or warmer than the present day, then the current warming phase can only be due to natural variation and no action is required to curb fossil fuel usage (see for example: [www.co2science.org](http://www.co2science.org)). The relative strengths, or otherwise, of this theory, and the causes of the current warming trend, are not the focus of this study. However, it is clear that there remains significant debate as to the extent of the impacts of the MWP and, it could be argued, that some aspects of this debate have been influenced by agendas secondary to scientific inquiry (Bradley *et al.*, 2003b).

There is also considerable debate regarding the extent of the Little Ice Age. Mann (2002) categorically states that “the notion of the Little Ice Age as a globally synchronous cold period has all but been dismissed” yet, according to Wanner *et al.* (2008) and Bütikofer (2007), the LIA is probably the only event in the last 6000 years to have occurred simultaneously in both hemispheres due to the combined effects of low Northern Hemisphere summer insolation, solar minima and volcanic forcing on the global climate system. The debate here is less about possible inferences that may or may not support individual arguments and more about the spatial synchronicity of events. To a large degree this debate revolves around the temporal scale at which the LIA is viewed. At a decadal level there is certainly significant hemispherical and regional variation in the timing of the onset of the LIA and of the coldest periods within the LIA (e.g., Bradley and Jones, 1993; Haston and Michaelsen, 1997; Jones and Mann, 2004; Moberg *et al.*, 2005). When viewed at a centennial scale however, there is a surfeit of evidence to conclude that, globally, the LIA was a period of significant climatic change (e.g., Mayewski *et al.*, 2004) and, in many places, represented the coldest period of the Holocene (Wanner *et al.*, 2008).

However, as is the case with the MWP, there are doubts surrounding the suitability of applying the term 'Little Ice Age' to records of climatic change that occur in wide ranging locations. Mayewski *et al.* (2004:244) specifically note that they refrain from using such terminology as 'Little Ice Age' as they feel it is "geographically or temporally restrictive". Jones and Briffa (2001:7) are also critical of the use of the term as it is "suggestive of an epoch colder than today ... [which is] likely to be inappropriate in many parts of the world". Ogilvie and Jónsson (2001) outline the history of the use of the term and summarise the issues surrounding its usage. They identify three main schools of thought and provide examples of the proponents of these. The first, as expressed by Lamb (1977), is the view that the LIA was a time of cold temperatures, not only in Europe, but in most parts of the world. The second, as proposed by Grove (1988), is that the LIA should be used only in regard to glaciation, not temperature. The third viewpoint is that the term is completely misleading and that there was no 'Little Ice Age' at all. Here, Ogilvie and Jónsson (2001) cite Landsbergis (1985:62) who states that ... "it would be far better to expurgate the scientifically misleading ... term 'Little Ice Age', because of the fact that the interval so designated was not uniformly cold in space or time".

It is clear then, that confusion exists surrounding the use of the term 'Little Ice Age', what it implies in terms of climatic change and what it is believed to represent. In a similar manner, as outlined above, the same can be said for the Medieval Warm Period. One possible reason for this is that no suitable definition exists of the temporal boundaries of these periods. In both instances, a lack of spatial coherence in the timing of the onset and cessation of the periods means it is difficult to definitively state that the findings of an individual study are representative of the impact of the MWP or the LIA. This is contrary to other, admittedly longer and higher amplitude, episodes, such as the Younger Drays, where chromosomes have been defined (Bork *et al.*, 1998).

In light of the above, the following discussion of centennial-scale climate change in the Northern and Southern Hemispheres is therefore undertaken with an acknowledgement that the terms LIA and MWP are not necessarily the most appropriate terms for the periods that they are argued to span.

### 2.3.2 The Northern Hemisphere

Of the studies examining climate of the last 2000 years, the clear majority emanate from the Northern Hemisphere (Jansen *et al.*, 2007). There are several factors contributing to this, including, among others, a long documented history which can provide proxy evidence of climatic change (e.g., Tarand and Nordli, 2001). The presence of such a number of studies can enable comparisons within and between regions, though the sheer volume of studies can also provide confounding interpretations due to the differing proxies, the relative responsiveness of proxies, the spatial and temporal limitations of proxies, differing dating resolution and standard dating errors (Jones *et al.*, 1998).

The classic depiction of climatic change in the Northern Hemisphere over the past two millennia is one of warm to cool cyclic change, which (using the commonly applied terminology) is typically represented as a Roman warm period (RWP), followed by a dark ages cool period (DACP), then a medieval warm period (MWP) followed by the little ice age (LIA). Such a sequence of events is evident in some individual studies from Iceland (Sicre *et al.*, 2008), Ireland (McDermott *et al.*, 1999), Alaska (Hu *et al.*, 2001; Loso *et al.*, 2006) and China (Ji *et al.*, 2005) among others (see Ljungqvist, 2009 for example), though it is not so clearly represented in composite Northern Hemisphere climate reconstructions (Briffa, 2000; Jones and Mann, 2004; Mann and Jones, 2003; Mann *et al.*, 2008, Moberg *et al.*, 2005). Indeed, across much of the Northern Hemisphere, climate reconstructions of the first millennium show highly disparate results.

For example, there is evidence of increasing summer precipitation in Finland (Helama *et al.*, 2009) and increasing temperatures in Fennoscandia (Seppä and Birks, 2001; 2002) around 800 AD, though these are not reflected in records from Sweden (Linderholm and Gunnarson, 2005) or Iceland (Axford *et al.*, 2008; Sicre *et al.*, 2008). Comparisons between two peat bogs in Ireland, and one in northern England, show a generally wet first millennium in all sites, though the English site indicates a prolonged prominent dry period between *ca.* 200 – 800 AD, not evident in the Irish sites (Barber *et al.*, 2003). Similarly, comparisons between six highly resolved records of diatom-inferred palaeosalinity from the Great Plains of North America indicate little coherences during the first millennium (Laird *et al.*, 2003) even though lakes from the region have been shown to respond in a coherent manner to climate forcing, in terms of epilimnetic temperatures at least (Benson *et al.*, 2000).



The prevalence of such intra-hemispheric disparities during the first millennium, possibly due to the RWP and DACP being of smaller magnitude than the MWP and LIA, makes inter-regional comparisons difficult. From the turn of the first millennium onwards, a slightly more coherent pattern emerges, though regional variability is still evident.

From Europe, Davis *et al.* (2003) used data from ~500 pollen records to reconstruct Holocene summer, winter and annual temperature variability. They divided the continent into six regions; northeast (NE), northwest (NW), central east (CE), central west (CW), southeast (SE) and southwest (SW), with results indicating that regional variability existed in all three reconstructions. Winter temperature reconstructions (the model with the highest cross-validated  $r^2$  correlation coefficient between observed and pollen-inferred climate values;  $r^2 = 0.83$ ) indicated that around the turn of the first millennium, temperature anomalies (compared to the 1890 AD reconstruction) were negative in the NW and NE regions, positive in CW and SE regions and negligible in the SW and CE regions. Annual temperature reconstructions ( $r^2 = 0.80$ ) of this time period also indicate notable variability, with a strongly positive anomaly in the CE region of approximately 1°C, a slight negative anomaly in the SW (~ -0.5°C) and negligible anomalies evident in all other regions including the northern regions. Also from northern Europe, Helama *et al.* (2009) find the existence of a ‘megadrought’ at the turn of the first millennium. Here, the authors examined more than 550 tree-ring records from Finland to reconstruct precipitation and found a significantly arid period from around 930 – 1220 AD which, they propose, was primarily caused by the influence of ENSO on the North Atlantic Oscillation. While this is not evident in the temperature reconstructions of Davis *et al.* (2003), the timing of this drought corresponds with a period of glacial retreat across much of Europe (Grove and Switsur, 1994) and elevated temperatures in the Alps (Mangini *et al.*, 2005).

In North America, there is broad-scale, tree-ring based evidence of a hydrological anomaly around 1000 AD (e.g., Graumlich, 1993; Stahle and Cleaveland, 1994; Cook *et al.*, 2004; Edwards *et al.*, 2008) which is replicated in numerous studies using other proxies (e.g., Luckman, 1994; Stine, 1994; Laird *et al.*, 1996a, 1996b, 1998; Li *et al.*, 2000; Hu *et al.*, 2001; Loso *et al.*, 2006; Meko *et al.*, 2007) and has been linked to the collapse of Mayan civilisation (Hodell *et al.*, 1995). Here too, the underlying cause of

the drought has been attributed to ENSO variation (Cook *et al.*, 2004, 2007; Rein *et al.*, 2004; Herweijer *et al.*, 2006, 2007; Seager *et al.*, 2007), which Seager *et al.* (2007) identify as being a persistent *La Niña*-like state with a warm Atlantic Ocean. They expand this hypothesis to suggest that a persistently positive North Atlantic Oscillation (NAO) would also lead to a warming of the Sargasso Sea. This, in turn, would explain the presence of medieval drought in parts of the Mediterranean (Bar-Matthews *et al.*, 1998) and northern Africa (Kondrashov *et al.*, 2005), and would lead to a wet northern Europe. While the findings of Abrantes *et al.* (2003) support the notion of a positive NAO, the assertion of a wet northern Europe contradicts the findings of Helama *et al.* (2009) outlined above, who found persistent drought conditions in Finland during this time. When combined with the findings of Verschuren *et al.* (2000) and Stager *et al.* (2003), who identify solar variability, rather than ENSO, as the cause of identified low lake-levels in equatorial Africa during medieval times, a claim also supported by Stuiver *et al.* (1997) and Bard and Frank (2006), it becomes clear that the primary driving mechanism behind the presence of such widespread hydrological anomalies is yet to be fully understood.

Despite this, the evidence from the Northern Hemisphere clearly indicates the presence of a widespread, extreme climatic anomaly around the turn of the first millennium. This took the form of a temperature and/or hydrological anomaly and, while the boundaries of this event differ from site to site, a shift away from this climate regime is generally noted by *ca.* 1400 AD, marking the beginning of the Little Ice Age.

As with the MWP, the signature of the LIA is widespread across the Northern Hemisphere. Temperature estimates suggest a gradual cooling trend after *ca.* 1400 AD (Mann *et al.*, 2003, 2008; Moberg *et al.*, 2005), reaching a minimum around 1850 AD, though the timing of this does vary (Guiot *et al.*, 2005; Nesje and Dahl, 2003). Generally identified as a period of glacial expansion across Europe (e.g., Grove, 1988), a shift away from the previously warm/dry climate in North America during the LIA is also noted (e.g., Haston and Michaelsen, 1997; Hu *et al.*, 2001; Wiles *et al.*, 1999), possibly driven by a warming of sea-surface temperatures in the eastern Pacific (Graham *et al.*, 2007).

The onset of the LIA, and the coldest periods within it, differed both spatially and temporally (Nesje and Dahle, 2003). Glacial advances occurred in Iceland (Caseldine

and Stötter, 1993), the European Alps (Holzhauser *et al.*, 2005), Scandinavia (Matthews *et al.*, 2000; Winkler, 2003; Bakker *et al.*, 2005), Alaska (Wiles *et al.*, 1999) and the Canadian Rockies (Luckmann, 2000), though the timing of glacial advances and retreats generally differed within, and between, regions. In most of Europe, summer temperatures were mild, with the exception of particularly cold periods in the seventeenth and early-nineteenth centuries (Bradley and Jones, 1993), while winter temperatures generally decreased (Davis *et al.*, 2003). The duration of sea-ice around Iceland increased (Ogilvie, 1984) and the River Thames periodically froze over in London (Fagan, 2000; Mann, 2002; Jones and Mann, 2004). Sea surface temperatures off Iceland were notably cooler (Andrews *et al.*, 2001; Sicre *et al.*, 2008), as was the Sargasso Sea further south (Keigwin, 1996).

In the Mediterranean region, an increase in freshwater diatoms in a marine core off Portugal suggests increased river discharge and, subsequently, higher precipitation, during the LIA (Abrantes *et al.*, 2003). Similarly, evidence of catastrophic flooding has been found off the Spanish coast (Vallve and Martin-Vide, 1998) and in the Rhône Delta in Mediterranean France (Arnaud-Fassetta and Provansal, 1999) around the same time. Similar evidence exists in other areas in the Mediterranean, such as Italy, Crete (Grove, 2001) and Tunisia (Marquer *et al.*, 2008), suggesting the influence of a strongly negative NAO.

Across North America, patterns of climate variability differed. In the southwest, cool summer temperatures and increased winter precipitation existed in the Sierra Nevada region of California (Graumlich, 1993), though there is intra-regional variation in precipitation anomalies evident (Haston and Michaelsen, 1997). In the central region of the continent, around the Northern Great Plains, there is evidence of significant variability. A study of six lakes (three in southern Canada, three in northern USA) seems to indicate a west to east pattern of variability with the western lakes being generally saline from *ca.* 1400 – 1800 AD, while the eastern lakes were generally fresh (Laird *et al.*, 2003). Previous studies have also suggested that the Northern Great Plains region may have been more arid during the LIA (Fritz *et al.*, 1994). Sea surface temperatures in the eastern Pacific were generally warm (Graham *et al.*, 2007), consistent with a prevailing *El Niño* state (de Batist *et al.*, 2008).

Across central Asia, there are examples of regional variation at decadal scales. For instance, the 1650s were exceptionally cold across Korea and eastern China, though were a period of relative warmth in Japan (Bradley and Jones, 1993). At centennial scales, Ji *et al.* (2005) reconstructed Asian monsoon intensity from central China and find that the period 1200 – 1600 AD was generally dry. This is corroborated by a study of aeolian dust deposition from northeastern China, which finds a peak in minerogenic clastic content of sediments around the same time (Chu *et al.*, 2009), and from a speleothem oxygen isotope study from Dongge Cave in southern China, which concurrently indicates a weakened Asian monsoon (Wang *et al.*, 2005).

The underlying cause of the LIA is still unclear. Broecker (2000) suggests that fluctuations in deepwater formation between the Atlantic and the Southern Ocean may be a factor, though bases this hypothesis on what he admits is mostly circumstantial evidence of fluctuating deepwater formation in the larger Dansgaard-Oeschger cycles. In contrast, Mauquoy *et al.* (2002), Stuiver *et al.* (1997) and Bard and Frank (2006) all suggest that solar variability is the driving force behind the climatic changes of the LIA. As is the case with the MWP then, there is no definitive agreement on the cause of the LIA. Indeed Wanner *et al.* (2008) suggest that perhaps no single mechanism is responsible for the LIA, rather a combination of events, including low Northern Hemisphere orbital forcing, low solar activity and an increase in the number of major volcanic events, may be the most likely cause.

### 2.3.3 The Southern Hemisphere

As discussed above, there is still debate as to whether the LIA and MWP were global events (e.g., Hughes and Diaz, 1994; Soon and Baliunas, 2003). To some degree, this debate is hindered by a lack of high-resolution palaeoclimatic studies from the Southern Hemisphere (Bradley and Jones, 1993; Jones *et al.*, 1998; Mann *et al.*, 1998, 2003, 2008). Even so, sufficient records exist to enable a general depiction of centennial-scale climatic change over the hemisphere to be presented. As with the discussion above of the Northern Hemisphere, the focus here will primarily be concerned with the period from the late first millennium onwards, as this contains the two most prominent climatic shifts of the last 2000 years.

In South America, a composite of Andean ice core records indicates periods of warm summer temperatures from 300 – 500 AD and 1100 – 1300 AD, followed by a period of cool summer temperatures from 1400 – 1900 AD, attaining a minimum over the 17th century (Thompson *et al.*, 2006). This is in general agreement, at centennial scales, with the record of glacial advances in Bolivia (Rabatel *et al.*, 2008), Peru (Jomelli *et al.*, 2008) and, further south, in northern Patagonia (Villalba, 1994), which all attained maximum advances, of varying extent, in the 17th century. In contrast, at least one glacier in northern Patagonia attained a maximum in the late 19th century (Araneda *et al.*, 2008), possibly reflecting the effects of local climate influences. During the 20th century however, all the glaciers discussed have retreated. Lake-levels in southern Patagonia were low from 1230 – 1410 AD and consistently high from 1500 – 1900 AD (Haberzettl *et al.*, 2005), which generally agrees with tree-ring based temperature and rainfall reconstructions from northern Patagonia and central Chile (Villalba, 1994). Further south in Tierra del Fuego, the dry period may have occurred earlier, around 960 – 1020 AD, though the authors concede, in this instance, that chronological control is poor for much of this record (Mauquoy *et al.*, 2004). From the east of the continent, there appears to be a contrasting pattern to that of the west. A pollen-based climate reconstruction from southern Brazil indicates that the period around 900 AD was particularly wet while the period from 1520 to 1770 AD was warm, at least in the highland regions (Behling *et al.*, 2004). A record of mangrove dynamics from the Amazon delta seems to corroborate this, with evidence of a lack of inundation apparent between 1130 and 1510 AD, and again from 1560 AD to the 19th Century (Cohen *et al.*, 2005).

The climate history from Africa is more complex. While there is an apparent synchronicity in the occurrence of several centennial-scale climate events in equatorial east Africa over the late Holocene (Russell *et al.*, 2007), including an arid phase around the turn of the first millennium (Russell and Johnson, 2005; Stager *et al.*, 2003; Verschuren *et al.*, 2000), this correspondence breaks down between *ca.* 1500 – 1800 AD. During this time, there is an apparent east-west gradient in rainfall variability with sites in the eastern equatorial region showing elevated rainfall and lake-levels, while sites in the western portion of equatorial east Africa indicate periods of drought (Russell *et al.*, 2007). It is postulated that the cause of this variation may be a combination of the influences of ENSO, a shift in the position of the ITCZ and coupled changes in the high-latitudes (Russell and Johnson, 2007).

Rainfall and temperature reconstructions from a speleothem in southern Africa demonstrate a very cool-dry period between approximately 800 and 1000 AD followed by a generally warm-wet, though variable, climate from *ca.* 1000 – 1500 AD (Lee-Thorpe *et al.*, 2001). After ~1500 AD, there is a dramatic and continual decline in temperatures and rainfall, attaining a minimum around 1800 AD. While these results appear to be similar to those from sites in the western portion of east Africa described above, the timing of the events differs at centennial-scales, suggesting a separate climate influence over the regions. This may take the form of the Indian Ocean Dipole (IOD; Saji *et al.*, 1999), an ENSO-like structure of fluctuating warm and cool sea-surface temperatures (SSTs) between the east and west Indian Ocean, which has been shown to significantly influence rainfall over southern Africa (Birkett *et al.*, 1999; Saji and Yamagata, 2003; Marchant *et al.*, 2007). Some support for this is evident in an SST reconstruction from Madagascar (Zinke *et al.*, 2004), in which the coolest SSTs correspond with the South African drought episode. While the degree to which the IOD is independent of ENSO is still debated (Saji *et al.*, 1999; Behera and Yamagata, 2003; Zinke *et al.*, 2004; Ihara *et al.*, 2008), the effect of the IOD on rainfall in south-eastern Australia has recently been more clearly defined (Meyers *et al.*, 2007; Ummenhofer *et al.*, 2009) and, in recent decades at least, there has been an opposing relationship between rainfall in eastern Africa and rainfall in the portion of the Australasian region influenced by the IOD (D'Arrigo *et al.*, 2008).

Due to its relatively recent discovery (Saji *et al.*, 1999), little is known of the palaeo-history of the IOD and, historically, there has been more of a focus on identifying past ENSO variability. Moy *et al.* (2002) identified significant *El Niño* events from sediments in Laguna Pallcacocha in the Ecuadorian Andes in which periods of increased alluvial deposition corresponds to the heavier rainfall events associated with moderate to strong *El Niño* episodes. Their results indicate that over the last two thousand years, there has been marked variability in the occurrence of significant *El Niño* events at centennial scales, while at a millennial scale, there has been a conspicuous increase of such events during last two millennia (Figure 2.4).

NOTE:

This figure is included on page 30 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 2.4: Reconstruction of significant El Niño events from Laguna Pallcacocha, Ecuador over the Holocene and the last 2,000 years. Top panel from Moy et al. (2002), bottom panel from data available at: <http://www.ngdc.noaa.gov/paleo/pubs/moy2002>.*

In the more recent portion of the Moy *et al.* (2002) record, there are a number of clear centennial-scale episodes of fluctuating *El Niño* intensity. Of particular interest is the period from *ca.* 800 – 1000 AD, which corresponds with several of the ‘medieval’ drought episodes in South America discussed above, and the reduced frequency and severity of drought events after *ca.* 1400 AD, which falls within the general timing of the LIA. In a synthesis of records examining past *El Niño* variability, Markgraf and Diaz (2000) present evidence from tropical South American ice cores to suggest that this period was dominated by a *La Niña*-type mode of the southern oscillation. If this was the case, it can be expected that the LIA in the western Pacific region would take the form of a generally warm and wet period. There is clear evidence to support this. From a palaeolimnological study in Java, Crausbay *et al.* (2006) find evidence of an “ENSO regime shift” after 1650 AD with a reduced frequency of drought events. Similarly, from a multi-proxy reconstruction of ENSO events since 1525 AD, Gergis and Fowler (2009:378) found a “prominent decrease in ENSO activity coincides with the duration of the LIA” and “the height of extreme *La Niña* activity is seen from the

16th to the mid 17th centuries.” Isotopic evidence from corals in Vanuatu show warm and wet conditions prevailed prior to *ca.* 1860 AD (Quinn *et al.*, 1993), and corresponding results are seen from the Great Barrier Reef (Hendy *et al.*, 2002a) and Rorotonga (Linsley *et al.*, 2000). The discontinuous Palmyra coral record also indicates comparatively warmer SSTs during the 17th century, though, in this case, the portions of the record that fall within the MWP – from 938 – 961 AD and 1149 – 1220 AD – temperatures were cooler (Cobb *et al.*, 2003). The timing of these events is roughly synchronous with periods of increased *El Niño* frequency in Ecuador (Moy *et al.*, 2002) and further indicate that, not only was the LIA warm in the western Pacific region due to a prevalence of *La Niña* events, the MWP was cool due to a prevalence of *El Niño* events.

Corroborating these findings with records from the Australian mainland, a region where ENSO dominates the climate system (Allen *et al.*, 1996; Nicholls, 1988), is difficult as millennium-length high-resolution studies are rare. However, a low-resolution record of aeolian dust deposition from the east of the continent does provide some support by indicating that wet conditions prevailed from around 1500 to 1750 AD, while dry conditions existed between *ca.* 900 and 1400 AD (Stanley and DeDeckker, 2002). Similarly, between *ca.* 900 and 1250 AD, Holdaway *et al.* (2002) found a lack of fire hearths from the “arid margins” of Australia and suggest that this indicates an abandonment of the region by indigenous populations due to a lack of permanent water sources. Primarily based on changes in sediment composition from Lake Keilambete in western Victoria, Mooney (1997) suggests a period of temperature variability existed between *ca.* 850 and 1200 AD, though, due to a low sedimentation rate above this period, there is little data from which to infer LIA climates. From Lake Euramoo in the tropical northern east coast, Haberle (2005) identified an increase in macroscopic charcoal particles between *ca.* 800 and 1400 AD, followed by a rapid reduction to very low charcoal levels between *ca.* 1400 and 1850 AD, also corroborating the proposed *El Niño* (MWP) to *La Niña* (LIA) shift.

In the south of the continent, temperature estimates from a tree-ring study from western Tasmania indicate a generally warm period between 900 and 1500 AD, followed by a reduction of multi-decadal variability between *ca.* 1500 and 1900, with a slight cooling compared to the preceding period (Cook *et al.*, 2000). However, this reconstruction is more closely aligned with Southern Ocean SSTs than ENSO and, for the period of the



LIA, may be reflecting an increase in westerly wind patterns that has been identified by Shulmeister *et al.* (2004). Temperature reconstructions from New Zealand are similar to those from Tasmania with a warm 13th century, though a more noticeable cooling is evident in the 17th century (Wilson *et al.*, 1979; Williams *et al.*, 2004) which corresponds to a period of glacial advances (Winkler, 2003, 2004) and an increased southwesterly circulation (Shulmeister *et al.* 2004).

Further south in Antarctica, Masson-Delmotte *et al.* (2004) examined an ice-core from Dome C and, using water stable isotopes, reconstructed temperatures of the past 5000 years from both the site of deposition (Dome C) and the moisture source (the subantarctic Indian Ocean). Their results indicate that, at centennial-scales, both site and source show a warming around the turn of the first millennium, followed by a cooling phase centred around 400 yrs BP. In east Antarctica, Goodwin *et al.* (2004) identify a weakly positive Southern Annular Mode, with reduced variability, between *ca.* 1550 – 1850 AD, indicating a poleward displacement of the mid-latitude westerlies over the region. From a marine core off the Antarctic Peninsula, Domack and Mayewski (1999) identified periods of increased productivity, indicative of elevated temperatures, centred around 1000 and 1300 AD. After 1300 AD, there was a dramatic decrease in productivity, which was sustained to the top of their record (*ca.* 1700 AD).

From the nearby South Shetland Islands, Serrano and López-Martínez (2000) identify a period of glaciation during the LIA, suggesting a temperature minimum and an increase in permafrost and periglacial conditions. Also from west Antarctica, Kreutz *et al.* (1997) identify a rapid change in meridional circulation based on an increase of sodium in an ice-core from Siple Dome, which they precisely date to 1399 AD. The deposition of sea-salt at the ice-core site is driven by the depth of the Amundsen Sea low pressure system which, in turn, is driven by temperature changes in the Pacific Ocean at inter-annual time scales, and the latitude of the Antarctic low-pressure belt at longer time-scales (Kreutz *et al.*, 1997). Interestingly, the timing and magnitude of this change in atmospheric circulation is reflected in the Greenland ice record, lending weight to the contention of global-scale climatic change during this period.

#### 2.3.4 Conclusions

It is clear that the stereotypical sequence of a ‘Medieval Warm Period’, followed by a ‘Little Ice Age’, is not applicable to many parts of the world. This is perhaps most notable in the western Pacific region, which appears to reflect the opposite of what is inferred by the terms. Indeed, in light of the above discussion, the use of terminology that solely focuses on temperature seems inappropriate given the variable nature of climatic responses. Even so, it is clearly apparent that, across the majority of the globe, climatic anomalies of some sort were centred around the turn of the first millennium and in the second half of the second millennium. In many regions the identified anomalies have been linked to ENSO or solar variability, though the underlying mechanisms driving centennial-scale climate change at such a scale have not yet been clearly identified.

In addition, it is also evident that, despite an increasing number of studies originating in the Southern Hemisphere over recent decades, some regions remain relatively understudied in terms of short-term (one to two millennia), fine-resolution studies that are able to identify climatic change at decadal to centennial-scale. Among these regions is Australia, which is the focus of the following discussion.

#### **2.4 High-resolution palaeo-archives: an Australian perspective**

The Australian continent lies within the field of influence of major atmospheric and oceanic climate systems such as ENSO, the Pacific Decadal Oscillation (PDO), the IOD, the Southern Annular Mode (SAM), the Inter-Tropical Convergence Zone (ITCZ) and the Indo-Pacific Warm Pool (IPWP) (Turney *et al.*, 2006). It is, therefore, ideally positioned to provide information regarding the variability and interactions within and between, these systems over time. In addition, as the continent contains regional climates ranging from tropical to temperate, and alpine to hyper-arid, there is the potential to examine the influence that historical fluctuations of these systems have on varying environments. However, as Europeans only colonised Australia in 1788 there is not a long cultural history from which to draw various forms of documentary evidence that could provide proxies for changing environmental conditions. In addition, instrumental climate records rarely extend beyond the late 19th century and, consequently, only provide a very brief insight into short-term, climatic variability. Therefore, there is a need use palaeo-archives in order to extend the climate record and

gain a greater understanding of natural climatic variability and the changing influence of oceanic and atmospheric climate systems. Several potential sources of such archives exist, though each has its own limitations. These are discussed below in the context of their potential application in Australia.

#### 2.4.1 Tree-rings

Tree-rings are a valuable source of high-resolution, precisely dated records that can, under the right conditions, provide data at up to seasonal resolution. In areas where long-lived species exist, or there are preserved sub-fossil remains of long-dead trees, records can be extended many millennia (Briffa, 2000). The main limitation of tree-ring studies is that not all tree species are suitable and the location of suitable species is usually limited to cooler regions where tree growth is slow and longevity is high (Briffa, 2000).

In Australia, despite several tree species indicating a potential for long-term studies (e.g., Heinrich and Banks, 2005), the only lengthy records have originated from studies of *Lagarostrobos franklinii* (Huon pine) on the west coast of Tasmania (Cook *et al.*, 1991, 1992, 2000; Buckley *et al.*, 1997). The longest of these extends to 1600 BC (Cook *et al.*, 2000) and provides detailed insight into short- and long-term temperature variability in western Tasmania and relationships with broad scale, sea-surface temperature changes. To date, this is the longest, continual, fine-resolution, climate reconstruction from anywhere in Australia.

#### 2.4.2 Ice-cores

Due to a lack of permanent ice fields, it is not possible to obtain ice-core records from Australia. However, some distant records from Antarctica have demonstrated correlations with climate systems that directly influence certain regions of Australia, such as the Southern Annual Mode (Goodwin *et al.*, 2004) and Southern Ocean temperature (Masson-Delmotte *et al.*, 2004). Other studies have reconstructed variables such as atmospheric circulation patterns over Antarctica (Kreutz *et al.*, 1997) that may have a more indirect influence on Australian climate. The potential to obtain very high resolution proxy evidence of climate systems affecting Australia is an exciting prospect for reconstructing Australian palaeoclimates.

### 2.4.3 Speleothems

Measurements of stable isotopes and trace element content of speleothems can provide highly detailed records of past climatic conditions (e.g., Lee-Thorp *et al.*, 2001) at seasonal to annual resolution. Records from Australia have been correlated against rainfall (Treble *et al.*, 2003, 2005) and ENSO (McDonald *et al.*, 2004) events and have provided evidence of environmental change throughout periods of the Holocene (Goede *et al.*, 1996), the last glacial (Goede, 1994) and the mid-Pleistocene (Desmarchelier *et al.*, 2000). At this point in time though, there has not been a continuous record developed for the late Holocene period as there has in other parts of the world (e.g., Bar-Matthews *et al.*, 1998; Lee-Thorp *et al.*, 2001; Williams *et al.*, 2004), or a combination of records to provide such. In this sense, speleothem-based climate reconstructions in Australia are, at this stage, of limited utility to investigations of the last two millennia.

### 2.4.4 Coral

As is evident from the discussion above, coral records can provide highly detailed records of changing oceanic conditions at seasonal or annual resolution, which can be linked with major climatic forces such as ENSO. In addition, the annual banding of corals enables the development of precise chronologies. As such, corals can provide a very powerful tool for reconstructing climatic variability over extended time-spans. The record of Hendy *et al.* (2002a) from the Great Barrier Reef provides evidence of sea-surface temperatures and sea-surface salinity variability over the past 420 years and provides valuable insight into ENSO variability over that period. Other coral records have also been correlated with discharge events from major river systems in the northern tropics of Australia, also providing insight into the past ENSO variability and the Australian monsoon (McCulloch *et al.*, 1994; Isdale *et al.*, 1998). The obvious limitation of coral records is that they are restricted in their spatial distribution to tropical and sub-tropical, shallow marine environments. In addition, a review of the relevant literature reveals that most coral records are of multi-century duration, with continuous records that extend beyond a millennium uncommon.

#### 2.4.5 Lake sediments

As sediments are deposited in lake environments, numerous isotopic, geochemical or biological proxies of environmental change are also laid down (Battarbee, 2000). The utility of these proxies to high temporal-resolution studies is partly dependant upon the nature of those proxies, though mostly dependant upon the rate at which sediment is deposited (Lotter and Sturm, 1994) and the degree to which it is disturbed after deposition (Larsen and MacDonald, 1993). Where suitable conditions exist though, lake sediments can provide very long and continuous records of past environments and offer great potential for reconstructing past climates (Battarbee, 2000). Some lakes, primarily those located in much colder climates than what the majority of Australia experiences, can provide annual or even seasonally varved sedimentary layers, the thickness and composition of which can be related to the climate (e.g., Loso *et al.*, 2006). Such sediments present an ideal opportunity for very high-resolution studies (e.g., Lotter and Birks, 1997; Lotter *et al.*, 1997b), though unfortunately, such highly-resolved sedimentary sequences have not been identified in Australia and, indeed, Australian lakes generally exhibit low sedimentation rates (Harle *et al.*, 2005), at least in the period prior to landscape clearance by European settlers.

Despite this, and in light of the discussion above, lakes offer one of very few potential options for continuous, high-resolution, continental studies from mainland Australia. As is the case with the other sources of palaeo-archives discussed above, lakes are also limited in their spatial distribution with, for example, no permanent lakes in the arid interior of the continent, excepting small, sparsely distributed mound springs (Mudd, 2000). However, where precipitation is more reliable and suitable morphometric features exist, lakes can be a common feature of the landscape. Western Victoria, in the southeast of the mainland, is one such area and consequently has become one of the most intensively studied regions of the country (Mooney, 1997).

The generally low sedimentation rates of Australian lakes (Harle *et al.*, 2005) may be a contributing factor as to why so few high-resolution palaeolimnological records have originated from Australia. To date, the only short-term studies that have been undertaken from western Victorian lakes are those by Mooney (1997), who examined a 2,000 year record using multi-proxy analysis, and Radke (2000) whose ostracod-based, Holocene climate reconstruction demonstrated a higher sampling resolution over the period of the record covering the last millennia. Neither record was of a sub-decadal

resolution though, primarily due to the short length of the cores (60 cm and 66 cm respectively) which precluded such temporal resolution. Recently, there has been an increased focus on retrieving highly resolved records from lakes in western Victoria (Peter Gell pers. comm.), while some sites in the northern tropics may also provide the potential for high-resolution analysis (e.g., Haberle, 2005; Haberle *et al.*, 2006).

Another aspect of high-resolution climate studies that requires consideration is the type of proxies that are to be examined as different lake-based proxies demonstrate different response times to climatic forcing. For example, terrestrial pollen, in particular that from trees, exhibits a slow response to a change in climate, with several decades often elapsing between the onset of a climate change, the demise of species in the affected region and, in particular, the colonisation and maturation of species that are more tolerant to the new climate regime (Green *et al.*, 1988). In comparison, due to their short life span, diatom species assemblages can change rapidly in response to changing lake conditions (Dixit *et al.*, 1992). An example of this is presented in a multi-proxy study from Lake Euramoo in the northern wet tropics of Australia (Haberle *et al.*, 2006). In a reflection of natural variability (Haberle *et al.*, 2006), diatoms species assemblages underwent a significant change around 1700 AD, though fluctuations in pollen, chironomids, charcoal and inorganic content of the sediment, show no evidence of this. In this sense, proxies that respond rapidly to climate-driven environmental change, and have a narrow window of tolerance to changing conditions, are more valuable to highly-resolved records. This is especially the case when seeking to identify high-frequency, climate processes that operate at sub-decadal time-scales (e.g., ENSO).

In terms of the biological record, there are several proxies that are suited to highly-resolved studies, each with their own advantages and disadvantages. In addition to diatoms, the most commonly employed indicators are ostracods and chironomids, though others include pigments, chrysophytes, cladocera, seeds, spores and sponge spicules (Armstrong and Brasier, 2005). Most of these indirectly reflect climatic change and are, instead, reflecting the nature of the lake's response to climatic change. The speed and nature of a lake's response to climatic change itself varies between lakes depending on such variables as the size and depth of the lake, exposure to wind and solar radiation, the size of the lake catchment, the lake's stratification regime, the ratio of littoral zone to open water and the nature of inflows and outflows (Mason *et al.*, 1994; Battarbee, 2000; Cohen, 2003). However, the nature of the lake's response is also

dependant on the nature of the climatic forcing. An increase in solar radiation, for example, can decrease water levels through evaporation, but can also increase the lake water temperature and induce stratification, thereby reducing the availability of nutrients and changing the chemical properties of the water itself (Cohen, 2003). Similarly, an increase in precipitation can raise water depth, deliver sediment and nutrients to the lake, affect turbidity and light penetration and alter hydrochemical parameters. Fundamentally then, the degree to which the proxies reflect climatic change is directly related to the degree to which the lake is climatically sensitive, which in turn is related to the morphometry of the lake (Larsen and MacDonald, 1993; Mason *et al.*, 1994; Battarbee, 2000). Appropriate site selection, and choice of proxies, is therefore of paramount importance (Fritz, 2008).

The process of site selection for this study is outlined in Chapter 4, while the primary proxy employed was diatom analysis. Principally, diatoms were chosen because they have proved to be powerful tools in palaeolimnological research (e.g., Fritz *et al.*, 1991; Dixit *et al.*, 1992; Battarbee, 2000) including a Holocene-length climate reconstruction from the study region (Gell *et al.*, 1994). Crucially, as outlined above, due to their short life-span and ecological sensitivity (Fritz *et al.*, 1991), diatom species' turnover occurs rapidly in response to changing conditions which thereby enables high-resolution analysis of short-term variability.

## **2.5 Diatoms as indicators of past climates**

Diatoms are microscopic algae of the class Bacillariophyceae. They are unicellular with siliceous cell walls, consisting of two valves that, when joined together, form a frustule (Stoermer and Smol, 1999). The morphological intricacy of the valves forms the basis of their taxonomy. Diatoms are abundant, and virtually ubiquitous, in aquatic environments and readily preserve as fossils in the majority of aquatic ecosystems (Reid *et al.*, 1995). Diatoms can be quickly identified to species level and beyond (Round *et al.*, 1990) and numerous taxonomic texts are available to guide identification (e.g., Foged, 1978; Krammer and Lange-Bertalot, 1986, 1988, 1991a, 1991b; Sonneman *et al.*, 2000). It is unclear how many diatom species exist, though Mann and Droop (1996) suggest that it may be in the order of 10,000. Due to their ubiquity, the very large number of ecologically sensitive species and a long history of taxonomic study (Mann,

1994), diatoms have become one of the most widely utilised palaeolimnological indicators (Moser *et al.*, 1996).

Under ideal conditions, diatoms frequently occur in very large numbers and even a small sediment sample of 0.1 – 0.2 cm<sup>3</sup> can yield in excess of 10<sup>6</sup> valves and often consist of more than 50 species. The sheer volume and diversity of valves and species, in conjunction with the ease of collecting multiple samples and aforementioned environmental sensitivity, makes diatoms ideally suited for statistical analysis. The application of statistical techniques, in the form of predictive models (transfer functions), has been undertaken for well over two decades (e.g., Gasse and Tekaiia, 1983). Transfer functions quantify the relationship between modern species assemblages and a suite of environmental variables from a calibration set of lakes/streams to determine species' tolerance and optima to the defined parameters. This relationship is then applied to down-core fossil diatom assemblages in order to quantitatively infer past lake conditions (discussed further below).

The use of diatoms in palaeoclimate studies is widespread, either as the sole proxy or as part of a multi-proxy approach, and significant insights into regional climates, and lakes' responses to climate, have been gained through their use (e.g., Gell *et al.*, 1994; Gasse *et al.*, 1997; Anderson, 2000; Cumming *et al.*, 2002; Ryves *et al.*, 2002; Tapia *et al.*, 2003; Verleyen *et al.*, 2003; Weckström *et al.*, 2006; Rühland *et al.*, 2008; Laird and Cumming, 2008). As outlined above, lakes respond differently to differing climate forces and, where this response is known, the utility of diatoms enables reconstructions of these varying responses. This commonly takes the form of quantitative reconstructions of a particular variable or variables which ideally fluctuate in accord with changing climates. However, knowledge of the autecology of species can also allow qualitative evidence of lake response to climatic change to be inferred. For example, where suitable lake morphometry exists (Fritz, 2008), changes in lake depth may be inferred by the fluctuating dominance of planktonic and benthic forms (Gasse *et al.*, 1989; Barker *et al.*, 1994a). Similarly, other variables such as temperature (Anderson, 2000), nutrient concentrations and ratios (e.g., Si:P; Kilham *et al.*, 1986) or light and turbulence requirements (Wang *et al.*, 2008), may reflect lake response to climatic forcing. However, diatom life-forms do not always correspond with a specified category (e.g., benthic *or* planktonic) and lake nutrient concentration and/or taphonomic



issues can affect the ratios between life-forms (Gasse *et al.*, 1997). Therefore, care needs to be exercised if undertaking such a qualitative approach.

In addition, Fritz (2008) cautions that, due to non-climatic factors, such as lake geomorphology and hydrology, the nature and magnitude of a lakes' response to climate forcing can be non-linear and non-stationary. This can make lake-based climate reconstructions complex. Similarly, Gasse *et al.* (1997) warn against drawing a simple linear relationship between lake palaeosalinity and climatic change in all instances, citing an example of the dissolution of salt crusts increasing conductivity during periods of rainfall. A similar example is provided by Telford *et al.* (1999) who found increasing diatom-inferred lake salinity during times where nearby lakes were full. In this instance, humid periods were associated with pulses of saline groundwater into the study lake, leading the diatom-inferred salinity reconstruction to indicate, on first principles, the opposite of the climate history. In this regard, hydrologically and morphometrically simple lakes, such as some closed basin crater lakes, provide an opportunity for more straightforward diatom-based palaeoclimatological investigations.

Closed basin lakes often show a clear response to changes in the balance between precipitation and evaporation, through fluctuations in the ionic chemistry of the lake water (Gasse *et al.*, 1997; Battarbee, 2000). The application of diatom-based inference models can reconstruct these fluctuations through time and, thereby, provide a surrogate record of climatic change, and the application of a diatom-salinity transfer function is a commonly used approach. Diatoms are highly sensitive to changes in ionic chemistry (Gasse *et al.*, 1997) and diatom-salinity transfer functions are often very robust, with  $r^2$  correlation coefficients between measured and diatom-inferred salinity, determined by cross validation, often above 0.85 (e.g., Wilson *et al.*, 1996; Ryves *et al.*, 2002; Yang *et al.*, 2003). Numerous such transfer functions have now been developed from regions of the world ranging from Greenland (Ryves *et al.*, 2002) to Antarctica (Verleyen *et al.*, 2003) and every continent in between. From the resulting substantial body of knowledge, the general strengths and limitations of diatom-salinity transfer functions can be determined.

In a comprehensive review specifically of diatom-salinity transfer functions, Gasse *et al.* (1997) outline a number of such factors which can confound the development of transfer functions, including diatom taphonomy and the dissolution of diatom valves. In

terms of taphonomy, Gasse *et al.* (1997) highlight the point that differing species inhabit different environments within the same lake. In this sense, when sampling for the development of a calibration set, a single sample from either the lake edge or the lake centre can not be considered to be an accurate representation of the lake's diatom community. Gasse *et al.* (1997) provide evidence of this from Lake Purrumbete in western Victoria, in which diatom samples were collected by sediment traps, phytoplankton sweeps and surface sediment samples from different locations within the lake over an annual hydrological cycle. Results demonstrated widely varying species assemblages within, and between, sampling locations at seasonal scales, despite relatively minor fluctuations in the lakes' hydrochemistry. This suggests that modern sites should be sampled several times over a year (or more) in order to obtain the most accurate representation of species/environment relationships. Though a contrasting viewpoint, offered by Anderson (1995:152), is that a single surface sediment sample, which may constitute between one and five years worth of diatom deposition, has the advantages of "smoothing out inter-annual variability in both biota and water chemistry". The problem with this approach though, is that the water chemistry sample, taken at the time of sediment sampling, is a discreet measurement of conditions at that point in time, not an average of several years of accumulated data, as is present in the diatom sample. When calibrating the species and environment data then, taxa that grew under previous chemical regimes will be associated with the present water chemistry. This is an obvious source of error, and highlights the need, where possible, to take several samples over the hydrological cycle of the lake.

Some of the  $1-r^2$  error in diatom-salinity transfer functions may arise through the effects of diatom dissolution (Barker, 1992; Ryves *et al.*, 2006). As mentioned above, dissolution of diatom valves was an issue raised by Gasse *et al.* (1997) and it, along with valve breakage, can prove highly problematic when analysing fossil species assemblages, as both these factors can reduce the number of complete valves that can be identified and counted. In the case of valve breakage, this can require the identification of broken valves which may bias the count towards species that are morphologically exceptional. Despite the inherent strength of diatom valves (Hamm *et al.*, 2003), breakage can occur in the sample preparation process by excessive agitation of the sample or centrifuging at high speed for long periods. In addition, Flower (1993) found that drying sediment, most notably carbonate-rich sediment, was a major cause of valve fragmentation. Under natural conditions, breakage can occur due to the action of wind

and waves or the reworking of sediment in shallow lake systems (Flower, 1993) Once the valves are fractured, they have a greater surface area and can therefore be more susceptible to chemical dissolution (Gasse *et al.*, 1997).

Dissolution of diatom valves can occur before, during and after deposition and is generally most prominent in saline (Ryves *et al.*, 2001, 2009) and strongly alkaline systems with a pH greater than 9.0 (Barker *et al.*, 1994b; Gasse *et al.*, 1997). In addition, the bioavailability of silica is also critical for the formation and preservation of diatom frustules and where silica saturation is low, for example where fluctuating Si-depleted groundwater levels interact with sediment pore water (Gasse *et al.*, 1997), valve dissolution can occur. The rate of sedimentation may also affect dissolution rates because, in systems with slow sedimentation rates, deposited fossil diatoms are subjected to the effects of pore water chemistry for longer periods than in systems with more rapid rates of sediment deposition (Ryves, 1994).

In addition to lake chemistry, diatom species composition can influence rates of dissolution. Preferential dissolution can occur in lightly silicified valves and/or in valves that have a high surface area to volume ratio (Gasse *et al.*, 1997). In terms of the development of calibration sets, this can significantly bias the data towards more robust species (Ryves *et al.*, 2001; 2006). Several studies have examined the rate of dissolution and the susceptibility of various species under varying chemical conditions (e.g., Barker, 1992; Flower, 1993; Barker *et al.*, 1994b; Ryves *et al.*, 2001). This has resulted in the development of dissolution indices for the more robust species (Battarbee, 1988; Barker, 1992; Ryves *et al.*, 2001, 2009) in an attempt to quantify the degree of preservation. The application of such indices may increase the accuracy of palaeosalinity reconstructions from impacted sites through a reduction in model error (Ryves *et al.*, 2009). As Gasse *et al.* (1997:555) state, “this is not a trivial problem as diatom dissolution is a complex, salt-specific process.”

Apart from the possibility of diatom dissolution at higher salinities, other issues also arise when undertaking reconstructions from saline systems. At higher salinities, diatoms can show decreased precision in salinity reconstruction (e.g., Fritz, 1990) as the taxa most frequently occurring in such waters have a broad salinity tolerance (Gasse *et al.*, 1997). As such, lakes above the polysaline range (20 – 30 g/L *sensu* Gasse *et al.*, 1987) can demonstrate decreased species diversity (Gell, 1995), which in itself can

decrease the predictive power of quantitative reconstructions (Birks, 1994). This observation was a key consideration in selecting fossil study sites with salinity less than 20 g/L (see Section 4.11) and the resulting transfer function was specifically developed for application to the fossil assemblages of such oligosaline sites. However, where the transfer function is to have a broader application to lakes of wider range of past and present salinities, a longer salinity gradient in the modern dataset is required and ideally, the “training set should ... be designed to represent the maximal range of modern environmental variation as possible for the biogeographical area of interest” (Birks, 1998:309).

Indeed, this is usually the case, as the majority of diatom-salinity transfer functions are developed from calibration sets with long salinity gradients. A notable exception is Ryves *et al.* (2002), whose transfer function was tailored specifically for use in West Greenland lakes where the diatom flora is of a subsaline, athalassic type, thus making the application of transfer functions developed elsewhere inappropriate. Their modern calibration set consisted of lakes that ranged in conductivity from 24 – 4072  $\mu\text{S}/\text{cm}$  (average 721  $\mu\text{S}/\text{cm}$ ) and, despite the short gradient, species assemblages demonstrated a clear response to conductivity with community turnover evident along the gradient. The model was also robust, with a high jack-knifed  $r^2$  (0.88) between measured and inferred conductivity and low predictive errors, while the application to a fossil record produced results consistent with the climate history of the region (Ryves *et al.*, 2002).

While developing a transfer function specifically for application in restricted circumstances may represent a resource-intensive method, the Ryves *et al.* (2002) study demonstrates the benefits of such an approach where other, previously developed transfer functions, can not be utilised. However, the risk in undertaking reconstructions from such short environmental gradients is that, due to periods of more severe climates for instance, lake conditions in the past may exceed the boundaries of the calibration set gradient. In this case, the accuracy of reconstructions of such periods may suffer due to the lack of modern analogues in the fossil data.

The issue of poor modern analogues is recognised as a potential problem for reconstructing salinity (and other variables) from fossil data (Birks, 1998). In this instance, species in the fossil record are either absent from the modern calibration dataset, or occur in such low numbers that an accurate determination of their tolerance

and optima to the variable of interest can not be ascertained. Such an issue may arise in regions where modern lake conditions have been heavily impacted by anthropogenic disturbance such as catchment clearance, agricultural runoff, eutrophication or acidification. Reconstructing pre-disturbance conditions in such situations can be problematic (Davies *et al.*, 2002). A potential solution lies in the sampling of more lakes, however, this is dependent upon the presence of unimpacted lakes. In many regions, the extent of disturbance may preclude this approach. The merging of regional datasets, as undertaken under the auspices of the CASPIA project (Juggins *et al.*, 1994) and the European Diatom Database (Juggins, 2001) for instance, may provide the most plausible solution to this issue as “the increased biological diversity of the combined dataset will help in the search for modern analogues of fossil assemblages” (Juggins *et al.*, 1994:192).

In conclusion, the utility of diatoms makes them a powerful tool in lake-based palaeoclimate reconstructions, however, numerous issues can confound the interpretation of diatom-inferred climate changes, only some of which are reflected in measures of model performance. To a certain degree, some of these issues can be overcome through the addition of further proxies (Gasse *et al.*, 1997), by studying more than one site, or by comparison with climate studies from the same region. In the absence of such supporting data, a cautious approach must be taken during interpretation of the data. The homogenisation of several datasets from differing locations (Juggins *et al.*, 1994), may also overcome some of these issues though, as demonstrated by Ryves *et al.* (2002), tailoring a model for use in specific circumstances and/or locations, can also provide accurate, single-proxy, reconstructions when other options are unavailable.

## **2.6 Statistical techniques for determining diatom species-environment relationships**

Since the study of diatoms began in the 19th century, associations have been noted between the species observed and the chemistry of their habitats. A classic example from the 1930s is the seminal work of Hustedt (1937 – 1939) who examined around 650 samples from various habitats in the Indonesian islands of Java, Bali and Sumatra. Hustedt was principally interested in species’ response to pH and classified the most commonly encountered species into five categories based on their tolerance to pH.

Despite reviews and modifications (see Battarbee, 1984 for instance), derivatives of Hustedt's classification system are still commonly used (e.g., Tipping *et al.*, 2002; Rott *et al.*, 2006; Finkelstein and Gajewski, 2007; Rouf *et al.*, 2009).

By the 1970s, a large body of observational autecological data had developed that enabled qualitative reconstructions to be undertaken based on known ecological characteristics of taxa (Battarbee, 1984; Battarbee *et al.*, 2001a). However, concerns about lake acidification in the 1970s and 1980s “promoted [the] need for unambiguous empirical evidence” (Battarbee, 1984:193) of lake acidification processes. This, in combination with increased computing power, led to significant advances being made in the development of numerical techniques (Birks, 1998). Foremost amongst these was the ability to develop and apply transfer functions, thereby enabling quantitative reconstructions of environmental variables. This advancement altered the field of palaeolimnology to such a degree that, in 1998, Birks (1998:307) declared that “palaeolimnology has shifted emphases from being a predominantly qualitative, descriptive subject to being a quantitative, analytical science.”

Transfer functions are now a regular feature of palaeolimnological studies and have been developed for numerous microfossils, including chironomids (e.g., Larocque *et al.*, 2001), cladocera (Lotter *et al.*, 1997a), chrysophytes (Pla *et al.*, 2003), pollen (Cook and van der Kaars, 2006), phytoliths (Prebble *et al.*, 2002), testate amoebae (Wilmshurst *et al.*, 2003) and diatoms which, being most relevant to this project, are the focus of the following discussion. Diatom-transfer functions have been developed for several environmental variables, though the most common are pH, salinity/conductivity (often in conjunction with ionic concentration and/or ratios) and nutrients (total phosphorus). Early diatom-transfer functions used methods such as Principal Co-ordinates Analysis (PCA; e.g., Tudor 1973) or Factor Correspondence Analysis (Gasse *et al.*, 1987) to infer taxon – environment relationships, though now, the most commonly used approach employs Weighted Averaging (WA) based (ter Braak and Barendregt, 1986; ter Braak and van Dam, 1989) regression and calibration.

The underlying principle of WA regression and calibration is that biota respond to a changing environmental parameter in a unimodal fashion (a Gaussian response), displaying an optimum, and tolerance range, to the variable. This is generally the case in nature (Birks, 1998), though where the environmental gradient is short, it may be more

statistically appropriate to employ linear techniques (ter Braak and Prentice, 1988). Through gradient analysis, the length of the gradient, and therefore whether unimodal or linear techniques are most appropriate, is able to be determined. Ordination methods, such as Correspondence Analysis and Detrended Correspondence Analysis, are increasingly used techniques in gradient analysis (Palmer, 1993).

Correspondence Analysis (CA) is one of the most commonly used multivariate data analysis techniques in ecological studies (Legendre and Legendre, 1998). However, the use of CA can be problematic due to the underlying assumption of a unimodal species – environment response. When plotting the determined relationship in ordination space, non-linearities will therefore be presented as an arch (Legendre and Legendre, 1998). This makes the spacing of samples along the first axis potentially unrelated to the amount of change present, while also exaggerating the length of the second axis. In order to overcome this, Hill and Gauch (1980) developed the new ordination method of Detrended Correspondence Analysis (DCA), in which arched data is detrended and rescaled.

Detrending the data is undertaken by dividing axis 1 into an arbitrary number of segments and making the mean axis 2 scores, within each segment, equal to zero (Gauch, 1982; Legendre and Legendre, 1998). Rescaling shifts the position of samples along ordination axes to make the differentiation diversity constant. In so doing, the units of axis 1 are expressed as standard deviations of species turnover and represent the species' gradient length. The lower the gradient length, the more likely that species occur at both ends of the gradient. In general, a gradient length of less than two standard deviations indicates a linear response by the species to a theoretical gradient (Birks, 1998), while a gradient length of four and above indicates that very few, if any, species are shared by either ends of the gradient and is therefore indicative of a unimodal response of the species towards the environmental variable (ter Braak and Prentice, 1988).

Assuming the species display a unimodal response, the environmental variable that is exerting the greatest influence on species distribution can be identified using a method of direct gradient analysis. Most commonly, the technique used is Canonical Correspondence Analysis (CCA) (Palmer, 1993). Unlike CA or DCA, which only analyse site and species data, CCA incorporates environmental variables into the

analysis. In doing so, CCA directly relates the composition of species assemblages to measured environmental variables (Palmer, 1993), and plots species against environmental data in ordination space, identifying the measured environmental variables that explains most species variance. A further advantage of CCA is the ability to assess the significance of the influence that the environmental variables have over the distribution of diatom species through the use of Monte Carlo permutation tests (Palmer, 1993).

Having identified the dominant variable, weighted averaging techniques can be applied to quantify individual species' unimodal response to that variable (regression), and apply this known response to fossil data (calibration). In the case of diatom species response to salinity, this implies that a diatom species will be most abundant when salinity levels are at the optimum level for that species: the species' optimum. Above and below that salinity, the abundance of the species will, ideally, decrease. In the regression step of the process, a species optimum is calculated as the average of all the values of the environmental variable of interest across the dataset in which the species occurred, weighted by the relative abundance of the species. The species' tolerance is estimated as the weighted average standard deviation (Battarbee *et al.*, 2001b). The calibration step of the process uses the average of the abundance of the various taxa in the fossil sample, weighted by their respective optima (Battarbee *et al.*, 2001b).

Therefore, in WA, averages are taken in both the regression and calibration steps of the exercise. This can result in a reduction in the range of values of the environmental data and is corrected for in one of two ways. The first is by linear deshrinking regression, reported as WA(classical), and the second is through an inverse regression approach, reported as WA(inverse). In terms of results, the differences between these two methods are subtle, though classical regression performs better for reconstructions at the ends of environmental gradients (Battarbee *et al.*, 2001b).

Weighted averaging does have some limitations though. These fall into two categories; 'internal' limitations of the method itself, and 'external' limitations due to the quality of the original dataset. The former are summarised by Birks (1994:115) who states that the "major disadvantages are that WA regression is sensitive to the distribution of the environmental variable in the training set, ... that it considers each environmental variable separately, and that it disregards residual correlation among taxa". An



improvement of the WA approach was developed by ter Braak and Juggins (1993) in the form of Weighted Averaging – Partial Least Squares (WA-PLS) regression. The additional components available in this method “utilize the residual structure in the species data to improve the species parameters (‘optima’) in the final weighted averaging predictor (ter Braak and Juggins, 1993:485). ter Braak and Juggins claim WA-PLS can reduce prediction error by 70% in datasets with low noise, though in noisy datasets, the reduction is much less. With the development of computer programs such as C2 (Juggins, 2007), it is a simple matter to develop models using WA and WA-PLS and identify the best performing technique for an individual training set. A further ‘internal’ limitation is that WA assumes all species exhibit a unimodal response to the environmental variable, yet some species may show a sigmoid increasing/decreasing, a skewed unimodal or a linear response (Racca *et al.*, 2001). Artificial Neural Networks (ANNs) are capable of incorporating and modelling these various responses, though comparisons between the methods suggest that their predictive power is similar to WA-PLS (Racca *et al.*, 2001).

‘External’ limitations to WA relate to the quality and internal consistency of the original dataset (Birks, 1998). Foremost amongst these is the distribution of an environmental variable across the dataset. If the distribution of sites is skewed towards one end of the variable’s range, the predicted optima of taxa may also be skewed in that direction. When variables show a skewed distribution, transforming the data by such means as log-transformation for example, may rectify the problem. A further limitation of WA transfer function revolves around the issue of diatom provenance where factors, such as habitat availability and taxon ubiquity, can skew reconstructions of the variable of interest (e.g., Sayer, 2001). Likewise, Saros and Fritz (2000) note that nutrient enrichment may expand the high end of a species’ salinity tolerance range which may also skew reconstructions of salinity. These factors highlight the importance of having sound ecological knowledge of the training set taxa because, as Birks (1994:109) states, “... any statistical regression and calibration will always produce some result, [therefore] it is essential to be able to evaluate the reliability of all reconstructions both statistically and ecologically.”

Despite these limitations, WA and WA-PLS regression and calibration techniques remain the most widely used methods for the development and application of transfer functions in palaeolimnology (Birks, 1998; Racca *et al.*, 2001). The ultimate utility of a

transfer function though, lies in its ability to accurately predict past conditions based on fossil diatom data. The measure of a transfer function's predictive power is reported as an  $r^2$  correlation coefficient, while the overall predictive error of the model is reported as the RMSE (Root Mean Squared Error). Validation of the model is undertaken both internally and externally. The most robust method of internal validation is through the use of leave-one-out cross validation, or jack-knifing. The jack-knifing process creates new datasets from the original training set by iteratively removing each of the sites in turn, then predicting the value of the environmental variable from that site using species optima and tolerances calculated from the remaining sites. The errors of prediction are tallied, with the overall model error expressed as the Root Mean Square Errors of Prediction (RMSEP), while the resulting  $r^2$  is often reported as 'jackknifed  $r^2$ ' or  $r^2_{\text{jack}}$  in order to differentiate it from the apparent  $r^2$  of the model that has not been cross-validated. Cross-validation is considered to provide a more realistic representation of the predictive capacity of the transfer function (ter Braak and Juggins, 1993).

'External' validation of the model can be directly achieved by comparing reconstructions against known environmental measurements, such as lake monitoring or climate data for instance (e.g., Fritz *et al.*, 1994; Laird *et al.*, 1996a, 1996b, 1998). Birks (1998:313) believes this is "[t]he most powerful means of validation", though it requires a precisely dated, and often finely-resolved, sediment sequence, which are difficult to obtain. If this method of validation is not possible, the reconstruction can be evaluated indirectly, using numerical techniques (Birks, 1998). Several such techniques exist (Birks, 1998), each with the general aim of quantifying the degree of similarity, or distance, between diatom assemblages in the fossil data and those in the calibration set. Reconstructions from fossil samples that contain a high number of species that are well represented in the training set will naturally return lower sample-specific errors. In effect, these techniques indicate a degree of confidence that can be attributed to reconstructions from each individual fossil-sample reconstruction.

In conclusion, despite the issues that exist in relation to weighted averaging regression and calibration which need to be considered when developing a model, these techniques have proved pivotal in the advancement of palaeolimnology (Birks, 1998). However, interpretation of palaeoenvironmental reconstructions must also take into account other factors that are external to the statistical model. These include factors such as the geomorphology and hydrology of the study lake, which can influence the lake's

response to external forcing (Fritz, 2008), and lake morphometry (Gasse *et al.*, 1997; Fritz, 2008), resource requirements (Kilham *et al.*, 1986; Saros and Fritz, 2002), taphonomic issues (Gasse *et al.*, 1997) and interactions between environmental variables (Saros and Fritz, 2000), which can influence diatom species composition. In the case of diatom-salinity reconstructions in particular, these factors can combine to make palaeoclimate reconstructions difficult in some circumstances (Gasse *et al.*, 1997; Fritz, 2008).

## CHAPTER 3 – THE STUDY AREA AND STUDY SITE CHARACTERISTICS

### Introduction

The broad geographical area of western Victoria can be further subdivided into biogeographic regions (bioregions) based on their environmental features and ecological characteristics (Figure 3.1). The region is roughly bisected by the elevated topography of The Great Dividing Range (labelled as the Central Victorian Uplands in Figure 3.1) and the Grampians Ranges. To the south of these ranges lie the western plains in which both study lakes and the majority of the modern surface sediment sample lakes are located. The western plains can be further subdivided into the Glenelg Plains, The Volcanic Plains, the Warrnambool Plains and the Otway Plains.

NOTE:

This figure is included on page 51 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.1: Bioregions of western Victoria and study site locations. Redrawn from Victorian government webpage:  
<http://www.nre.vic.gov.au/plntanml/biodiversity/what.htm>*

### 3.1 Geology and Landform Evolution of the Western Plains

The Western Plains were formed during the rift between Australia and Antarctica roughly 60 to 110 million years ago (Mya) (Mutter *et al.*, 1985) and are overlain by Cretaceous sandstone and Tertiary limestone and marl (Price *et al.*, 1997) which are, in turn, overlain in many localities by a thin layer of lava flows. Volcanism was probably linked to tectonic movement over one or more hot-spots (Cohen *et al.*, 2008) and occurred in two main stages, with peaks of activity from the Palaeocene to the Eocene and again from the Pliocene to the Pleistocene (Wellman, 1974) or early Holocene (Joyce, 1988a). The former formed what is referred to as the 'Older Volcanics' and lie to the east of Melbourne, while the latter are referred to as the 'Newer Volcanics' and make up the Volcanic Plains Province which stretches to the west of Melbourne.

The Volcanic Plains cover approximately 15,000 km<sup>2</sup> and are marked by over 400 eruption points of late Tertiary to Holocene age (Joyce, 1988a). The newer volcanics can be further subdivided into two sub-provinces; the Central Highland (or Earlier New Volcanics) sub-province to the north of Melbourne aged 6 – 7 Mya and the Western Plains (or Newer New Volcanics) sub-province aged from 4.5 Mya to less than 20,000 years. The latter category includes Blue Lake in Mount Gambier, South Australia which was initially thought to have last erupted between 4000 – 4300 years ago (Joyce, 1988b), though further dating suggests this may be nearer to 28,000 years ago (Leaney *et al.*, 1995). While the 'earlier New Volcanics' are mostly heavily weathered, many of the younger volcanics, such as those from Mount Eccles, have had less weathering time and so have well preserved flow-features with little or no soil coverage (Head *et al.*, 1991).

Topographically, the western plains can be characterised as a generally low, undulating landscape with numerous volcanic cones scattered across its breadth. The volcanoes of the western plains can be divided into four types; lava volcanoes, which constitute almost half those in the region, though they are generally located in the central and eastern part of the Central Highland sub-province; scoria volcanoes, numbering almost 200, that are mainly located in the western part of the central highlands and southern part of the western plains; maars, which number around 40 and are mostly found on the southern edge of the western plains; and volcanic complexes, of which only a few exist and occur where two or more volcanoes were formed by eruption from a common

magma chamber (Joyce, 1988b). Many relic volcanoes now contain lakes or swamps. Some of these are in contact with, and maintained by, local groundwater (e.g., Tower Hill and Blue Lake; Leaney *et al.*, 1995) while others are closed systems (e.g., Lake Keilambete; Bowler and Hamada, 1971).

These lakes are often referred to as ‘Crater Lakes’, though this term is not entirely accurate. Timms (1992:62) states that “[t]rue crater lakes ... lie in former vents at summits of non-active volcanoes” and nominates Lake Surprise and Lake Mumblin (8 km from Lake Elingamite) as the only examples in the western plains. The other main category of lakes, that is of interest to this project, is maar lakes. Timms (1992:63) states that “[m]aars are formed by single volcanic explosions ... [often caused by] water that is explosively converted into steam on meeting hot lava ... [which] produce deep holes with little ejecta”. Examples in the study region include Lake Elingamite, Lake Keilambete, Lake Bullenmerri, Lake Purrumbete and, in South Australia, Blue Lake. The last three examples are the deepest lakes in the region at approximately 60 metres, 45 metres and 70 metres respectively (Timms, 1976; Leaney *et al.*, 1995).

### **3.2 Climate of Western Victoria**

The climate of western Victoria is generally characterised as being of a ‘Mediterranean’ type with hot dry summers and cool wet winters. Climate varies from very humid in the Otway Ranges to semi-arid in the Murray Mallee to the north. Both mean annual temperatures and rainfall deficit increase from south to north, only interrupted by the orographic influence of The Grampian Ranges in the centre of the region.

#### **3.2.1 Rainfall**

Across the western plains, mean annual rainfall varies between 600 mm and 1600 mm and, generally, the gradient decreases latitudinally with distance from the coast (Figure 3.2). The key orographic features of The Grampian Ranges, in the centre of the study region, and The Otway Ranges, in the south east of the region, significantly affect localised rainfall. The majority of rainfall occurs in the winter months of June/July/August, though significant rainfall events occasionally occur between March and October due to the influence of the northwest cloud bands which emanate from the tropical Indian Ocean and move southeast across the continent. The frequency of these

is intimately related to the Indian Ocean Dipole (Smith *et al.*, 2000; Ashok *et al.*, 2003), the Indian Ocean equivalent of ENSO. The primary influence on rainfall across the study region though is the prevailing westerly winds which bring moist maritime air masses into the region. The strength and persistence of the westerlies is, in turn, influenced by the Southern Annual Mode (SAM) oscillation (Australian Bureau of Meteorology (AusBOM); climate influences).

NOTE:

This figure is included on page 54 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.2: Mean annual rainfall in western Victoria. Isohyets in millimetres per year. Redrawn from Australian Bureau of Meteorology (AusBOM; climate maps)*

The SAM (also referred to as the Antarctic Oscillation) is a north/south oscillation of the mid-latitude westerly belt. A number of statistical methods have been used to define the strength of the SAM (e.g., Goodwin *et al.*, 2004; Hendon *et al.*, 2007) and while studies differ in their preferred definition, the underlying concept is based on

differences of varying – 500, 850, 1000 hPa – geopotential height anomalies between the mid-latitudes and the polar regions (Goodwin *et al.*, 2004). Periods of positive SAM are associated with a poleward contraction of the westerly belt and result in weaker than normal winds across southern Australia and higher surface temperatures. Conversely, in periods of negative SAM, the belt of westerly winds move northward, resulting in more storm systems affecting southern Australia and decreased surface temperatures (Hendon *et al.*, 2007). Each SAM event (positive or negative) can last up to 2 weeks with the timing between events generally in the range of a week to months (AusBOM; climate influences). Marshall (2003) showed that the SAM has demonstrated a trend towards more positive values over the last three decades (the limit of reliable data), which has, in turn, been associated with decreased winter rainfall and increased summer rainfall in the southeast of Australia (Hendon *et al.*, 2007). The interactions between the SAM and other climate processes is discussed further in Section 3.24.

Over a longer time period, it was noted by Gutteridge *et al.* (1980; cited in De Deckker and Williams, 1988) that the mean annual rainfall in the study region had actually increased after the mid-20<sup>th</sup> century. They demonstrated that, over the period of 1898 – 1945, mean annual rainfall was 738 mm, compared to 840 mm for the period of 1945 – 1980. While Gutteridge *et al.*'s (1980) data was from one weather station (Camperdown) only, a similar trend can be seen in rainfall anomaly data averaged across the state of Victoria, which reveals low average values until 1945, followed by increased values until 1980 (Figure 3.3). Indeed, Suppiah and Hennessy (1998) found that, over the period of 1910 – 1990, the number of dry days had decreased over southeastern Australia, while total rainfall and the number of heavy rainfall events had increased. Interestingly though, since the mid-1990s, and after the publication of the papers mentioned above, annual rainfall anomalies (from a 1961-1990 baseline) have shown a steady negative trend, with negative values recorded every year since 2001. In the annual climate statement for the year 2007, the Australian Bureau of Meteorology highlighted the severity of the current drought by noting that “southeast Australia has now missed out on the equivalent of an average year’s rainfall over the past 11 years” (AusBOM; climate statement 2007).



NOTE:

This figure is included on page 56 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.3: Victorian annual rainfall anomaly (base 1961-1990) with 11-year running average. (Base average 693.74 mm). Source: AusBOM; Vic rainfall*

### 3.2.2 Temperature and evaporation

The temperature gradient across the western plains is strongly latitudinal, increasing with distance from the coast. For example, in Terang (Latitude 38.24°S, elevation 131 m) the mean annual maximum temperature is 18.5 °C, compared to 21.5 °C for Horsham (Latitude 36.66°S, elevation 128 m) (AusBOM; climate data) approximately 180 km to the north. Similarly, mean annual evaporation increases from 1200 mm in the south to 1400 mm near Horsham (AusBOM; average rainfall), though generally only exceeds precipitation from October to April (Williams, 1981).

Mean annual temperature anomalies in Victoria show a steady increase since the 1950s (Figure 3.4). Indeed, since 1900, a total of 33 years have shown positive values (relative to the 1961 – 1990 baseline), with 25 of these years occurring since 1972 and positive values occurring continually since 1996, with 2007 being the hottest year on record. Similarly, mean annual sea surface temperature anomalies from south of the study region have shown a continual warming since the 1930s (Figure 3.5). Furthermore, by increasing evaporation, the recent rise in temperatures has increased the severity of recent droughts when compared to previous droughts with comparable rainfall (Nichols, 2004; Ummenhofer *et al.*, 2009).

NOTE:

This figure is included on page 57 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.4: Victorian Mean Annual Temperature Anomalies (base 1961-1990) with 11-year running average. (Base average 14.1°). Source: AusBOM; Vic temp*

NOTE:

This figure is included on page 57 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.5: Southern Region Annual Sea Surface Temperature Anomaly (base 1961-1990) with 11-year running average. (Base average 14.11°). Source: AusBOM; SST*

### 3.2.3 El Niño-Southern Oscillation (ENSO)

Rainfall variability across the majority of the Australian continent is intimately linked to the *El Niño*-Southern Oscillation (ENSO) (Nicholls, 1988). The most commonly used measure of ENSO variability, the Southern Oscillation Index (SOI), is calculated as the mean sea level pressure difference between Tahiti and Darwin. Results are usually presented in terms of a monthly, or annual, average (Figure 3.6). Sustained negative values of the SOI indicate an *El Niño* phase, while sustained positive values indicate a

*La Niña* phase. Generally, *La Niña* conditions are associated with higher than average rainfall across the eastern and northern regions of the continent, while the *El Niño* phase is associated with lower than average rainfall (Nicholls and Kariko, 1993).

NOTE:  
This figure is included on page 58 of the print copy of  
the thesis held in the University of Adelaide Library.

*Figure 3.6: Annual SOI averages 1876-2007 with 11-year running average.  
Source; AusBOM; SOI*

Typically, Australia experiences drought across the eastern two-thirds of the continent, including the study area, during prolonged *El Niño* phases (Wang and Hendon, 2007). However, the strength of the *El Niño* cycle does not necessarily correlate to the strength of the corresponding drought (Wang and Hendon, 2007). For example, the 2002/03 *El Niño*, was only considered to be ‘weak’ with average annual SOI values of -6.1, yet the corresponding impact was classed as ‘very strong’ and brought record temperatures, severe rainfall deficits and contributed to major bushfires in New South Wales, the Australian Capital Territory and Victoria. In comparison, the 1997/98 *El Niño* was considered ‘strong’ with average annual SOI values of -11, yet the corresponding impact was only classed as ‘weak’ (AusBOM; *El Niño*). The non-uniform nature of the *El Niño*-strength to impact-strength relationship is due to several factors, such as the modulating influence exerted over ENSO by other regional ocean-atmospheric phenomena. These include the Inter-Decadal Pacific Oscillation (Power *et al.*, 1999) and the Indian Ocean Dipole (Saji *et al.*, 1999), discussed further in Section 3.2.4 below, or the magnitude and location of the sea-surface temperature anomaly in the equatorial Pacific (Wang and Hendon, 2007).

Recent studies have found that the occurrence of *El Niño* events has increased over the latter half of the 20<sup>th</sup> century. Gergis and Fowler's (2006) multi-proxy reconstruction of extreme ENSO events since 1525 AD shows that 43% of extreme events occurred in the 20<sup>th</sup> century, while 30% of all events have occurred since 1940 AD, with a bias toward *El Niño*. In a study from Pacific Ocean coral (not included in Gergis and Fowler's, 2006 data), Urban *et al.* (2000) found a significant shift towards *El Niño* conditions occurred in 1976, and has persisted since. This finding is reflected in average annual SOI data from Australia (Figure 3.6) and is a condition which may be due to the influence of greenhouse warming (Timmermann *et al.*, 1999).

It is difficult to definitively state the impact that ENSO has across the study area, let alone at individual sites, though, in general, there is increased rainfall over the region during extended *La Niña* phases, while the opposite is true of extended *El Niño* phases. The degree to which these phases impact the study region though, is affected by the influence of other regional ocean-atmospheric circulation phenomena.

#### 3.2.4 Drivers of climatic variability in south-eastern Australia

Climate variability in south-eastern Australia is influenced by four main climate processes; the Indian Ocean dipole (IOD), the southern annular mean (SAM), the inter-decadal Pacific oscillation (IPO) and, as mentioned above, ENSO (Power *et al.*, 1999; Kiem and Franks, 2004; Gillett *et al.*, 2006; Murphy and Timbal, 2008). These processes operate on differing time scales and interactions between them can moderate or exacerbate their relative impacts on the climate of the study region (Power *et al.*, 1999; Meyers *et al.*, 2007; Pezza *et al.*, 2007; Murphy and Timbal, 2008; Ummenhofer *et al.*, 2009).

The SAM has been described as the leading mode of variability in the southern hemisphere (Pezza *et al.*, 2007) and, as discussed above, is a measure of the latitudinal change of the mid-latitude westerly winds. The winds are displaced southwards when the SAM index is positive, and northwards when they are negative (Thompson and Wallace, 2000). Using data from 1979 onwards, Hendon *et al.* (2007) compared daily rainfall data over Australia during the positive phase of the SAM index and found that during this phase, the southeast of the continent experienced decreased winter rainfall

and increased summer rainfall. Similar results are also reported by Gillett *et al.* (2006). Given the sampling resolution employed in the lake studies presented here, seasonal-scale variability can not be discerned from the record, however Thompson *et al.* (2000) found that the SAM has demonstrated a significant upward trend over the last 50 years which suggests a potential contributing factor to the increasing rainfall anomalies over Victoria since the 1960s (see Figure 3.3).

However, the SAM does not operate independently of other climate processes. L'Heureux and Thompson (2006:281) identified a positive correlation that exists between the SAM index and ENSO during the summer months and find that “~25% of the interannual variance of the SAM ... is described by temporal variability in the ENSO cycle”. They illustrate this by demonstrating that the largest warm ENSO events of 1982/83, 1991/92 and 1997/98 are all mirrored by decreases in the SAM index. This indicates that ENSO plays a role in modulating the SAM and, therefore, the impacts described above.

In comparison to the relatively recently discovered relationship between the SAM and ENSO, the impacts of ENSO on Australian rainfall have been known for sometime and are widely documented in the literature (e.g., Nicholls, 1988; Wang and Hendon, 2007). Research over the last decade has identified the impact of the Pacific decadal oscillation (PDO)<sup>2</sup> and the inter-decadal Pacific oscillation (IPO) on modulating the impacts of ENSO (Gershunov and Barnett, 1998; Power *et al.*, 1999; Salinger *et al.*, 2001; Kiem and Franks, 2004). In particular, the incidence of flood risk, and the severity of individual events, have been found to be greater during times of negative IPO and positive SOI (Kiem *et al.*, 2003; Micevski *et al.*, 2006). In specific relation to the study region, the southeast of the continent also receives higher than average rainfall during *La Niña* phases (Murphy and Timbal, 2008). However, this is more pronounced when

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<sup>2</sup> The PDO and the IPO indices are derived from different approaches yet they have been shown to be very similar with the main difference being that the IPO relates more the Pacific Ocean as a whole and operates on interdecadal timescales, whereas the PDO refers more to the North Pacific and operates on sub-decadal timescales (Folland *et al.*, 2002; Kiem and Franks, 2004). Both have been found to impact the Australian climate through interactions with ENSO (Power *et al.*, 1999; Salinger *et al.*, 2001) though whether one has a greater impact than the other has not been identified. This may, in part, be due to the relatively recent identification of both processes which has lead to confusion over the names. Indeed, Mantua and Hare (2002) claim that the IPO is simply another name for the PDO. Folland *et al.* (2002) however, state that the IPO “can be regarded as the quasi-symmetric Pacific-wide manifestation of the PDO”, while along similar lines, Salinger *et al.* (2001) argue that the IPO may be “the Pacific-wide manifestation of the PDO, excluding subdecadal timescales”. Therefore, when comparing records with sub-decadal variation the PDO is more relevant, whereas the IPO is more relevant to interdecadal comparisons.

the occurrence *La Niña* corresponds with times of increased sea surface temperature (SST) changes in the Indian Ocean off the coast of Indonesia, rather than in the Pacific Ocean as reflected by the IPO (Meyers *et al.*, 2007).

Relationships between SSTs in the Indian Ocean and Australian climate have been recognised for some time (Streten, 1981, 1983). Nicholls (1989) identified a correlation between Indian Ocean SST variability and the presence of rain bearing northwest cloud bands which stretch from the northwest of the continent to the study area in the southeast and are an important source of rainfall to the region (Murphy and Timbal, 2008). Nicholls (1989) also identified that the strength of these patterns were positively correlated with SSTs near Indonesia and negatively correlated with SSTs in the central Indian Ocean. The variability in Indian Ocean SSTs was further elucidated by Saji *et al.* (1999) who coined the term Indian Ocean Dipole (IOD) for the variation in SSTs of opposite polarity between eastern and central Indian Ocean. Saji *et al.* (1999:361) proposed an index to identify the strength and variability of the IOD over time, which is calculated as “the difference in SST anomaly between the tropical western Indian Ocean (50°E – 70°E, 10°S – 10°N) and the tropical south-eastern Indian Ocean (90°E – 110°E, 10°S – Equator)”. The index is positive when SSTs are cool at the eastern pole and warm in the western pole and negative when the reverse is true. Using instrumental data, Saji *et al.* (1999) reconstructed the IOD from 1958 and comparisons between the IOD and Pacific SSTs suggested that the dipole was independent of ENSO, though this is still debated (Behera and Yamagata, 2003; Zinke *et al.*, 2004; Ihara *et al.*, 2008).

Following the work of Nicholls (1989) and the formal definition of the IOD and development of the IOD index by Saji *et al.* (1999), further studies have examined the impact of the IOD on Australian rainfall (Smith *et al.*, 2000; Ashok *et al.*, 2003) and the relationship the IOD has, if any, to ENSO (Behera and Yamagata, 2003; Zinke *et al.*, 2004; Meyers *et al.*, 2007; Ihara *et al.*, 2008; Ummenhofer, 2008; Ummenhofer *et al.*, 2009). Of particular interest to the study region is the work of Meyers *et al.* (2007), who examined the SST data from the Pacific and Indian Oceans of all the years between 1877 and 1998 and formalised a classification system to determine if they were characterised by *El Niño*, *La Niña* or ‘no event’ and/or positive IOD, negative IOD or ‘no event’. These results are summarised in Table 3.1 and it is immediately apparent that the combinations of negative SOI/negative IOD and positive SOI/positive IOD rarely, if ever, occur. Of the remaining seven classifications, Meyers *et al.* (2007) found

a high probability of good rainfall in the southeast of Australia existed when negative IOD occurred alone and when negative IOD co-occurred in conjunction with *La Niña* (Figure 3.7). In an extension of these results, Ummenhofer *et al.* (2009) compared the relative importance of the IOD and ENSO to the severity of drought in south-eastern Australia. Using the classification system of Meyers *et al.* (2007) and comparing against the National Oceanic and Atmospheric Association's Palmer Drought Severity Index (PDSI), which also incorporates the effects of surface air temperatures, Ummenhofer *et al.* (2009) demonstrated that Indian Ocean SST variability was the primary cause of droughts over south-eastern Australia, and was, in fact, more of a determinant than ENSO.

*Table 3.1: Classification of the years between 1877 and 1998 according the results of Meyers et al. (2007). Values represent the number of years in that classification, while values in parentheses represent the number of years which have a high degree of certainty in the classification assigned to them. Values in blue indicate the characteristics required to induce a high probability of good rainfall in the southeast of Australia, values in red indicate the characteristics required to induce a high probability of drought.*

NOTE:

This table is included on page 62 of the print copy of the thesis held in the University of Adelaide Library.

NOTE:

This figure is included on page 63 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.7: The SST characteristics and resulting impacts on rainfall anomalies of various combinations of ENSO and IOD between 1877 and 1998, from Meyers et al. (2007). a. Composite averages of SST anomaly (°C) during June-November for years in parentheses in Table 3.1. b. Percentage of years with below median rainfall during June-November for years in parentheses in Table 3.1.*



### 3.3 Palaeoclimates of the Western Plains

The number and nature of, and ease of access to, the lakes in the western plains of Victoria make the region very attractive for palaeoecological studies and, as such, the region is one of the most intensively studied regions in Australia. Records cover time periods up to one million years old (Wagstaff *et al.*, 2001) and have been derived by a variety of means such as sedimentology (Bowler and Hamada, 1971; Bowler, 1981), ostracod analysis (De Deckker, 1982; Chivas *et al.*, 1986), palynology (Dodson, 1974, 1979; Edney *et al.*, 1990), faunal remains (De Deckker, 1982) and diatoms (Yezdani, 1970; Tudor, 1973; Gell *et al.*, 1994; Gell, 1998). The majority of sites studied lie within a 75 kilometre radius (Figure 3.8) and, as described above, within a similar climatic zone. By far the majority of records cover the Holocene and the very late Pleistocene, though sufficient records exist to provide a description of vegetation and, by extension, climatic change from the mid-Pleistocene onwards (e.g., Kershaw *et al.*, 2004a; 2004b).

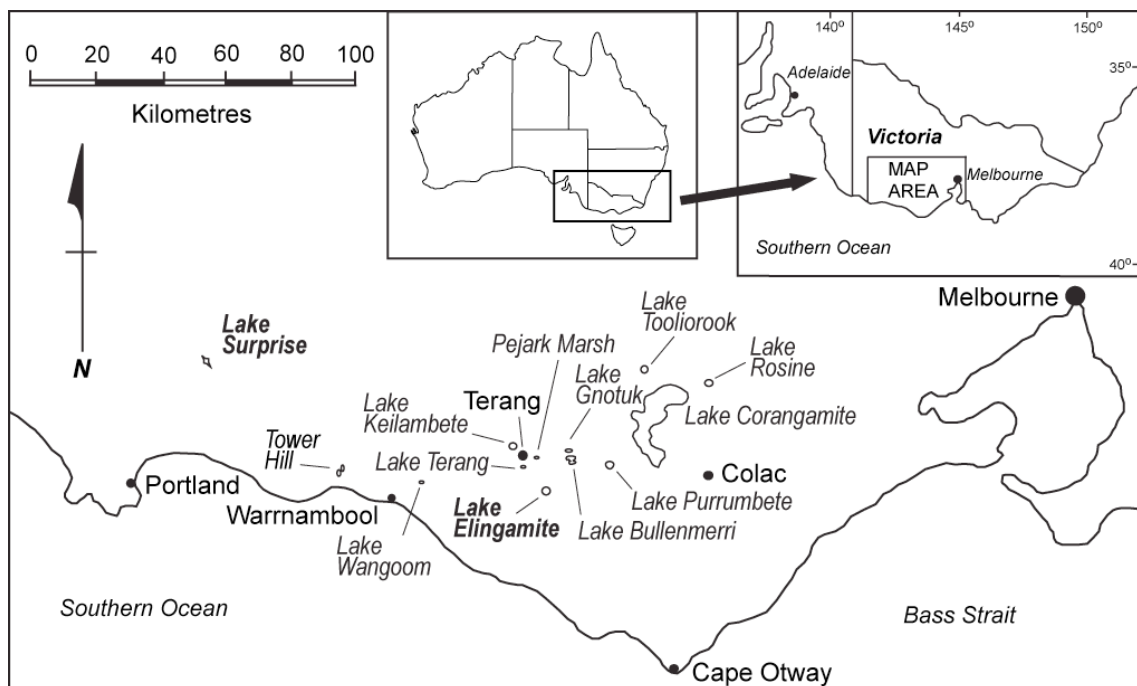


Figure 3.8: Map showing the location of study lakes and the location of various other lakes mentioned in the text

### 3.3.1 The Pleistocene

From the wealth of potential study sites in the western plains, there are presently only three sites that provide sedimentary records extending beyond 30,000 yr B.P.; Lake Terang, Lake Wangoom and Pejark Marsh. These are complemented by one further study from offshore. Of these, Pejark Marsh is the most ancient, dating from approximately 1,030,000 – 650,000 yr B.P. (Wagstaff *et al.*, 2001), followed by Lake Terang at approximately 425,000 yr B.P. (D’Costa, 1989; D’Costa and Kershaw, 1995; Urban *et al.*, 1996). These sites are less than four kilometres apart and their combined histories have been summarised by Kershaw *et al.* (2004b) who have also inferred mean annual rainfall from the palynological records evident. The next oldest record is from the Lake Wangoom study (Harle *et al.*, 1999) which is potentially a 200,000 year record (Harle *et al.*, 2002). The offshore record however, has a strong marine oxygen-isotope based chronology which dates the base of this record to 125,000 ka (Harle, 1997).

The early to mid-Pleistocene record from Pejark Marsh (Wagstaff *et al.*, 2001) implies a gradually drying climate, with the pollen history showing a change from a deep lake to increased swamp and scrub species. Kershaw *et al.* (2004b) suggest that rainfall at the time fluctuated from approximately 400 – 800 millimetres per annum. From the mid-Pleistocene onwards there was greater environmental variability, though perhaps a slightly wetter climate with a stable swamp community and peat development. The interglacial periods were characterised by an increase in precipitation to around 1250 mm/year as suggested by the presence of cool temperate rainforest taxa in the record (Kershaw *et al.*, 2004b).

The last glacial/interglacial cycle is captured in both the Lake Wangoom record (Harle *et al.*, 1999, 2002) and the offshore record of Harle (1997). Dating uncertainties confound the Lake Wangoom record, but Harle (1997) offers a very detailed interpretation from the offshore core with maximum precipitation attained at the height of the last interglacial with forest and rainforest widespread. Precipitation fluctuation was synchronous with changes in the marine oxygen-isotope record, slightly increasing through interstadial periods, but generally decreasing toward the last glacial period with *Eucalyptus* forest being replaced by a more open heath and mallee vegetation (Harle, 1997). The vegetation during the last glacial maximum (LGM) is interpreted as a grass and heath steppe, representing a cool dry climate.

While the sampling resolution of the offshore core E55-6 was relatively coarse (39 non-contiguous samples covering a total of 125,000 years), some corroboration of the interpretation, at least for the period leading up to the LGM, comes from Lake Leake (Dodson, 1975) and Wyrie Swamp (Dodson, 1977) in the south-east of South Australia. Records here indicate that eucalypt woodland and heath dominated from 50,000 yr B.P. to approximately 39,000 yr B.P. before being replaced by an open eucalypt woodland environment which persisted until the Holocene.

The height of the LGM occurred between 20,000 and 15,000 yr B.P., when the extent of aridity, temperature minima and continentality reached a peak (Dodson and Ono, 1997). It is therefore unsurprising that few records from the region fully cover this period, a fact which, in itself, indicates the degree of aridity. Those sites which do record the LGM include Lake Bullenmerri (Dodson, 1979), Northwest Crater at Tower Hill (D'Costa *et al.*, 1989) and Lake Surprise (Builth *et al.*, 2008). The vegetation of the western plains at the time is one of a semi-arid, grassland-steppe environment (Dodson and Ono, 1997; Builth *et al.*, 2008) that was virtually devoid of trees (Kershaw, 1998). An environmental setting such as this has no modern analogues in the Australian context, making rainfall projections problematic (Kershaw *et al.*, 2004b). However, Horton (1984, cited in Kershaw *et al.*, 2004b) estimates a mean annual rainfall of approximately 250 mm derived from the discovery of fossil Red Kangaroo, a semi-arid dwelling species, within the region. Temperature estimates of the time range from 4°C (Kershaw, 1998) and 6°C (Chappell, 1991) cooler than present day with higher wind speeds, increased seasonality, frosts and incursion of cold polar air as potential reasons for the lack of tree coverage over the region (Kershaw *et al.*, 2004b).

Between 15,000 yr B.P. and the Holocene period, temperature gradually increased and the vegetation evolved from a steppe-like environment to a more grassland environment. *Eucalyptus* species began to spread from the highland refugia where they had survived the LGM (Kershaw, 1998). The persistence of moderate levels of Chenopodiaceae pollen, in several records, is interpreted as indicative of a reduction of effective precipitation, caused by increasing temperatures, without a concurrent rise in precipitation (Kershaw *et al.*, 2004b). Evidence from Tower Hill (D'Costa *et al.*, 1989) and Lake Bullenmerri (Dodson, 1979) suggest this aridity continued until around 12,000 yr B.P., after which there was a significant climatic amelioration and an increase

in precipitation leading into the Pleistocene/Holocene boundary (Kershaw *et al.*, 2004b).

### 3.3.2 The Holocene

The generally wetter climate of the Holocene, being more amenable to the maintenance of sedimentary archives, is reflected by the number of records from the region which cover this time period. The record which perhaps best reflects the climatic fluctuations in the region over The Holocene period is the study of Bowler (1981) who used sediment analysis to reconstruct lake depth and, by inference, salinity at Lake Keilambete. These findings were supported by those of Chivas *et al.* (1986) from the same lake (Figure 3.9), while results from Lake Bullenmerri and Lake Gnotuk exhibit a similar pattern (DeDeckker, 1982). In general, these records indicate an increasingly humid environment over the first two millennia of The Holocene, followed by an extended period of around 4000 years in which effective moisture was high. The period between 4000 and 2000 yr B.P. was more arid, with climatic instability evident in the results of both the Bowler (1981) and Chivas *et al.* (1986) studies, while from 2000 yr B.P. approximately until European settlement, greater effective moisture is indicated.

Several authors have summarised the numerous Holocene-age studies from the region in order to provide a synthesis of vegetation and climatic change over the period, including Dodson and Ono (1997), Kershaw (1998), Dodson (2001) and Kershaw *et al.* (2004b). In general, open grassland and woodlands were gradually replaced by the expansion of *Casuarina* and *Eucalyptus* as temperatures and rainfall gradually increased between 10,000 and 7000 yr B.P.. High values of tree taxa between 7000 and 5000 yr B.P. suggest the maximum expansion of dense forest vegetation, though high percentages of *Pomaderris* also indicate greater tall open forests of which there are no modern analogues (Kershaw, 1998). Maximum effective precipitation was reached during this period due to higher rainfall and lower temperature extremes which, in turn, led to most lakes being full to overflowing at this time, as is evident in the Lake Keilambete record (Figure 3.9). After approximately 5000 yr B.P., lake levels gradually declined (e.g., Bowler, 1981; Chivas *et al.*, 1986) though despite the evidence of climatic instability evident in the Lake Keilambete records, the pollen records generally indicate little

vegetation change during this time, with the established vegetation cover remaining unchanged until the arrival of Europeans (Kershaw *et al.*, 2004b).

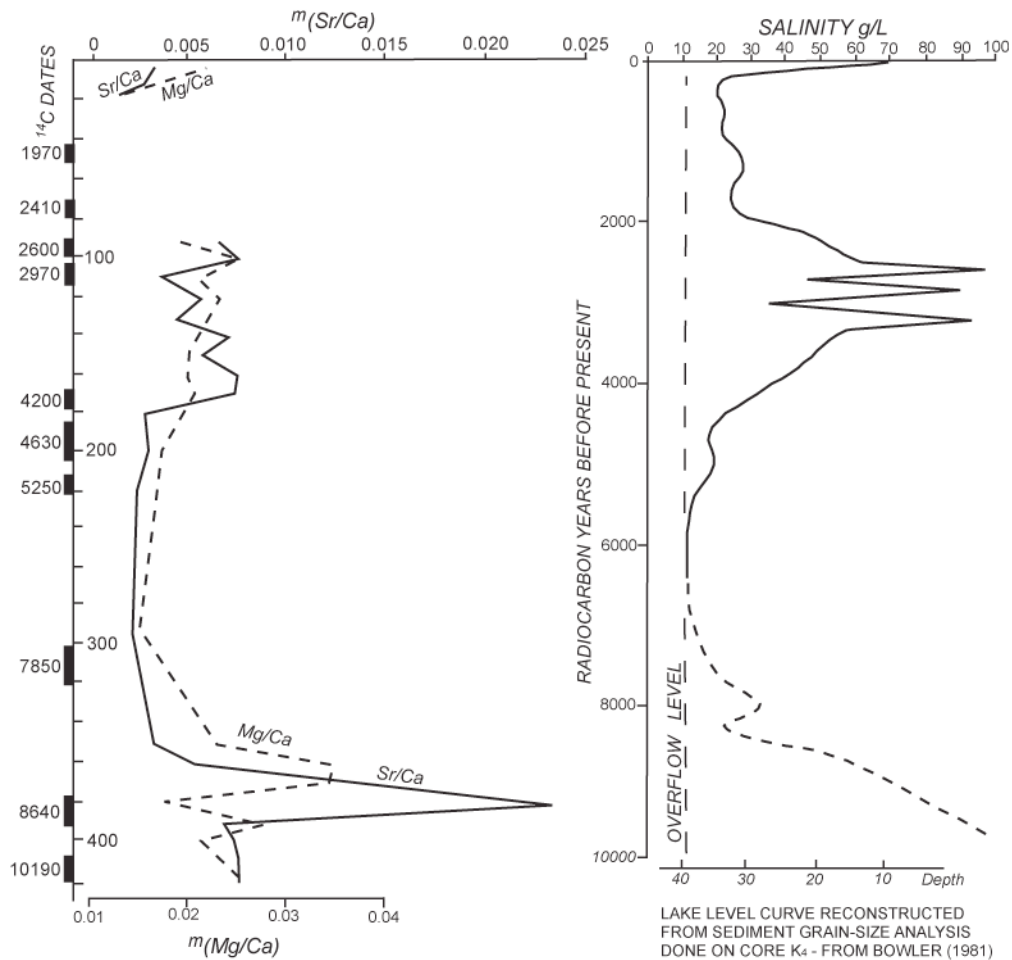


Figure 3.9: Holocene lake level and salinity reconstructions from Lake Keilambete derived from; (left) trace element analysis of ostracod shells (Chivas *et al.*, 1986) and (right), sediment analysis (Bowler, 1981). Redrawn from Chivas *et al.* (1986) and Bowler (1981).

Around 2000 years ago there was an increase in effective precipitation as lake levels increased, drowning shoreline trees at Lakes Bullenmerri (Gill, 1971) and Keilambete (Bowler and Hamada, 1971). This climate regime continued until around the 1840s or 1860s (Jones *et al.*, 1993; 2001), after which lake levels gradually declined across the region. In Lakes Keilambete and Bullenmerri, the re-emergence in the late 1960s of trees drowned approximately 1900  $^{14}\text{C}$  years ago suggests that the current climate is the driest for approximately 2000 years. This is probably due to a change in both precipitation and evaporation across the region (Jones *et al.*, 2001), though groundwater

abstraction has also been implicated in the recent decrease of some lake levels in the region (Timms, 2005).

### **3.4 The fossil study lakes**

The choice of study lakes is primarily influenced by the aims of the project and the methods with which those aims are to be achieved. While there is a wealth of potential study sites within the western plains region, not all are suitable to diatom-based, short-term, palaeoclimatic reconstructions. As such, a suite of criteria were defined to identify the lakes most suitable for this study. The rationale behind the criteria, and the process of elimination, which led to the selection of Lake Elingamite and Lake Surprise as the study lakes, are described in detail in Chapter 4.

Both lakes are located within the volcanic plains province and, while they lie approximately 100 km apart (see Figure 3.8), have similar latitude with Lake Surprise situated slightly to the north (Lake Elingamite: 38°21'S; Lake Surprise: 38°03'S). Data from the weather stations nearest to each site indicates that Lake Surprise annually receives an average of 17 mm of rain more than Lake Elingamite (798 mm compared to 781 mm). This difference may be a function of the length of the records though, as there was an increase in average annual rainfall across the region in the middle of last century (see Section 3.2.1 above) and rainfall data for the Heywood weather station (the nearest to Lake Surprise) has only been collected since 1949 compared to 1897 at Terang, the nearest weather station to Lake Elingamite (AusBOM compare). No evaporation data exist from either of these weather stations, though the Australian Bureau of Meteorology indicates the regional mean annual evaporation rate of the region to be in the range of 1200 – 1300 mm (AusBOM; Evap). In general, it is apparent that the lakes share a similar climatic regime and therefore, due to the distance between them, any palaeoarchives retrieved from them could, when viewed in tandem, reasonably be considered to be indicative of a regional pattern.

While the lakes share a similar climatic regime, they differ in almost every other aspect. Lake Surprise is an elongate crater lake, while Lake Elingamite is a circular maar, Lake Surprise is surrounded by native vegetation, whereas Lake Elingamite's catchment is cleared and, where Lake Surprise is narrow and protected by steep crater walls, Lake Elingamite is broad and largely unsheltered. Therefore, while the ambient rainfall and

temperature data are similar, the differential influence of wind, light and others factors are likely to yield contrasting moisture balance outcomes. Further individual characteristics are discussed in Sections 3.4.1 and 3.4.2.

#### 3.4.1 Lake Elingamite

The word “Elingamite” is an anglicised version of ‘Yelingamadj’ which was the name of the aboriginal clan who historically lived in the vicinity of the lake (Clark, 1990). The lake lies approximately eight kilometres southwest of the township of Cobden at latitude 38°21’S, longitude 143°00’E. This region was one of concentrated volcanic activity, with several other maars and craters nearby. Lake Keilambete, Lake Bullenmerri, Lake Gnotuk, Lake Purrumbete, Lake Terang, Cobrico Crater and Pejark Marsh are located within a 20 km radius (Figure 3.8). Lake Elingamite is a roughly circular maar, approximately 2.1 km in diameter, which is enclosed by a tuff ring of irregular height which is highest on the northern and eastern sides of the crater and lower and broader on the western side. An overflow point, leading to Elingamite Creek, lies on the north-western side of the crater (see Figure 3.10) and was approximately 11 metres above lake levels in 2006. It is likely that the lake overflowed around the turn of the 20th century (Timms, 1977; Mann, 1991). A disused quarry exists on the north-eastern side of the crater, part of which forms the road that provides access to the lake.

The tenure of the lake and foreshore area on the north-east shore is public land where a playground, picnic area, fishing clubhouse and boat ramp are located. The area within the crater has been largely cleared of overstorey vegetation, though over the last decade a significant revegetation project has been undertaken resulting in several thousand trees, mostly eucalypts, being planted around the western, northern and north-eastern shorelines (Ken Farquarson, pers. comm.). The majority of the remainder of the land inside the crater is privately owned and is principally used as pasture for the grazing of cattle. The surrounding landscape probably supported an open woodland environment prior to settlement though is now predominantly grasslands and pasture with an increase in land clearance occurring during the soldier settlement period post WWII (Mann, 1991).

NOTE:

This figure is included on page 71 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.10: Satellite image of Lake Elingamite, taken in 2006. Note the steep rim around the north-eastern edge of the crater and the distance from the boat ramp to open water. Dashed line represents the approximate size of the lake in 1977 (estimated from Timms, 1977). Image from GoogleEarth*

The lake has a reputation as a fine fishing location and is classified as a salmonoid fishery under the Victorian Inland Fisheries Strategy of the Department of Primary Industries. Each year the lake is re-stocked with fingerlings of non-native Brown Trout (*Salmo gairdneri*), Rainbow Trout (*Salmo trutta*) and, occasionally, Redfin (*Perca fluviatilis*). The last restocking took place in October 2007 and consisted of 500 Brown and 500 Rainbow Trout (Russell Smith, pers. comm.). Recently, fishing in the lake has been hampered by low water levels preventing the launching of boats.



NOTE:

This figure is included on page 72 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.11: Modelled lake levels for Lake Elingamite, from Mann (1991). Observations of authors mentioned in the text have been added and are represented by stars.*

Water levels have been gradually declining in the lake since the middle of last century (Figure 3.11). Hussainy (1969, cited in Timms, 1977) found the lake to be 12 metres deep in 1966, though Yezdani (1970) estimated it to only be “about 30 feet deep [ $\sim 9$  m]” in 1970. The most extensive study of the morphometry of the lake, undertaken by Timms (1977), measured a maximum lake depth of 9.2 metres. At the time of initial sampling for this project, in 2003, the lake had a maximum depth of 3.4 metres and observations, during visits to the region over the intervening years, have shown a continued reduction in lake levels (Figure 3.12), to the point where launching a boat from the boat ramp is no longer possible. Measurements from GoogleEarth satellite imagery indicate that the mean lake diameter in 2006 was approximately 1000 metres, in comparison to the 1930 metre mean lake diameter measured by Timms (1977). Modelling of lake levels between 1976 and 1990 (Roger Jones, unpublished data) suggest that the lake is responding to decreased moisture, however, water abstraction also appears to have played a role in declining lake-levels (Mann, 1991).



*Figure 3.12: Lake Elingamite in December 2003 (top) and March 2006 (bottom) illustrating decreasing water levels during the course of this project.*

Water abstraction for local crop irrigation occurred between 1964 and 1990 with an annual extraction entitlement of 399 megalitres (ML) (Mann, 1991). Local community agitation led to a government investigation into the cause of the declining lake levels. The outcome was the report of Mann (1991) who calculated that, under natural conditions (and 1990 rainfall and evaporation data), the total annual recharge into the lake, including the groundwater component, was approximately 170 ML, whereas the total discharge was calculated at 210 ML/year. The main discharge process was shown to be evaporation, which accounted for 170 ML/year, with the remaining 40 ML/year lost through aquifer recharge. Therefore, under a 1990 climate regime, the lake was found to have a naturally deficient water budget of approximately 40 ML/year and, as a result, Mann recommended the cessation of water abstraction.

The basin of the lake is gently funnel-shaped (see Figure 3.13), which is in contrast to other maars in the region which are all flat-bottomed (Timms, 1977). Timms (1977) attributes the irregular shape of the floor as probably being due to inward slumping from the crater rim, though there is no other evidence to support this hypothesis. The lake is in contact with the local aquifer (Figure 3.14), which contributes approximately 30 ML/year to the annual water budget, though this is countered by the approximately 40 ML/year that is lost from the lake via recharge back into the aquifer (Mann, 1991). The throughflow of groundwater is the most likely explanation for the continuing fresh nature of the lake (Table 3.2), despite an obvious decline in water levels. Reed swamps, mainly consisting of *Phragmites australis*, encircle the majority of the lake and

submerged aquatic macrophytes, including *Triglochin procera* in particular, occur in dense patches in the shallower water near the lake edge. The geochemistry of the lake water was determined by Maddox (1967) and Gell (1995) who both found the lake to be dominated by  $\text{Na}^+$ ,  $\text{Cl}^-$  and  $\text{Mg}^{2+}$ , and to a lesser extent  $\text{Ca}^{2+}$ .

NOTE:  
This figure is included on page 74 of the print copy of  
the thesis held in the University of Adelaide Library.

*Figure 3.13: Bathymetry of Lake Elingamite in 1977 with depth contours in metres. The greatest depth at the time of coring in 2003 was 3.4 metres. The dashed line represents the edge of the crater. Redrawn from Timms (1977).*

NOTE:  
This figure is included on page 74 of the print copy of  
the thesis held in the University of Adelaide Library.

*Figure 3.14: The interpreted hydrological regime of Lake Elingamite. Redrawn from Mann (1991)*

Table 3.2: Previous salinity and pH measurements from Lake Elingamite

Author	Salinity (g/L)	pH
Maddox (1967)	1.09	8.82
Hussainy (1969)*	1.05 – 1.4	Not measured
Yezdani (1970)	1.09	8.5 - 9
Tudor (1973)	1.1	7.5
Gell (1995)	1.89	9.3
<i>This study</i>	2.117	8.01

\* - cited in Williams (1981)

Yezdani (1970) retrieved a 130 cm sediment core from a “sandy beach area” on the northern shore of the lake which is no longer identifiable. No dating was carried out on the core, though *Pinus* pollen was found to a depth of 30 cm and diatoms were found in varying densities in the deeper parts of the core. A total of 40 diatom species were identified amongst the modern phytoplankton of the lake, the composition of which, when combined with other phytoplankton communities, led Yezdani (1970) to characterise Lake Elingamite as slightly to moderately eutrophic. Water analysis for this project does not support this assertion though, with total phosphorous concentrations being below the level of detection, with a value of < 0.1 mg/L (results of water chemistry analysis are provided in Section 5.4.2).

#### 3.4.2 Lake Surprise

Lake Surprise is one of only two ‘true’ crater lakes in western Victoria, the other being Lake Mumblin (Timms, 1975). The lake lies in the craters of Mount Eccles at latitude 38°04’S longitude 141°55’E and is situated within the 6120 hectare Mount Eccles National Park. Approximately 550 metres in length and 185 metres wide at its widest point, the lake is an irregular shape due to it occupying the contiguous remains of one large crater and two smaller ones (Timms, 1975). The gradient of the walls of the crater ranges from steep to vertical with a mean height of 49 m (Timms, 1975), offering considerable wind protection to the lake (Figure 3.15).

While no detailed hydrological study has been undertaken on the lake, the primary source of limnological information is the study of Timms (1975). He mapped the bathymetry of the lake (Figure 3.16) as well as examining water chemistry and other physical and chemical features. Timms found that the lake was dominated by Na<sup>+</sup> and

Cl<sup>-</sup> ions (as with many waters from south-east Australia (Williams, 1981)) and was slightly alkaline, ranging from a pH of 7.7 in winter to 8.5 in summer. Secchi depth ranged from 1.5 metres in summer (reduced by a “phytoplankton bloom”) to 8.2 metres in late spring, while salinity was reasonably constant ranging from 0.37 – 0.39 g/L. In spring and summer, the lake has a strong thermocline and stratifies with a sharp delineation between an oxygen saturated (120%) epilimnion and a completely de-oxygenated (0%) hypolimnion. At the time of coring (March, 2004), secchi depth was only 0.6 metres, pH measured 8.68, dissolved oxygen was 95.8% at the surface and electrical conductivity was 220  $\mu\text{S}/\text{cm}$ .

The lake is fringed by a band of swamp vegetation which is mostly concentrated at the shallower ‘ends’ of the lake (not visible in Figure 3.15). The surrounding vegetation is described by Builth *et al.* (2008:415) as “dominated by an open forest of *Eucalyptus viminalis* with a second storey of *Acacia melanoxylon* and under-storey of mainly Poaceae, Asteraceae, *Pteridium esculentum* and *Banksia marginata*”.



*Figure 3.15: Lake Surprise. Looking south from the rim of Mount Eccles crater and illustrating the height and gradient of the crater walls which form the immediate catchment of the lake (Image courtesy of Aline Philibert).*

NOTE:  
This figure is included on page 77 of the print copy of  
the thesis held in the University of Adelaide Library.

*Figure 3.16: Bathymetry of Lake Surprise. Depth in metres. Source: Timms (1975)*

The lake is the site of several ongoing palaeoecological studies based on an 18.5 m sediment core retrieved in 2004, which has a basal age of approximately 29,000 years, or an earlier 13.5 m core, retrieved in 2002. A further 110 cm frozen spade core of the uppermost sediments has also been obtained, which was utilised in this project as core LSFS (see Section 4.1.3.2). Pollen analysis has been undertaken on the long core (Builth *et al.*, 2008) and on core LSFS (Chris White, unpublished data). Results are illustrated in Figures 3.17 and 3.18. Diatom analysis has been undertaken on the 13.5 m core (Aline Philibert, unpublished data; Figure 3.19), while sedimentary mercury levels have been measured on the long core and core LSFS (Jacqui Hellings, unpublished data; Figure 3.20).

The pollen and diatom records demonstrate changes both within the lake and of the surrounding environment. Builth *et al.* (2008) assert that an increase in *Myriophyllum* species (specifically *M. muelleri*) and *Botryococcus* (not included on their diagram) during the LGM indicates that the lake was shallow and probably brackish at this time, while the transition from Asteraceae to Casurinaceae to Eucalyptus is indicative of a shift towards more humid climates from the late-Pleistocene to early-Holocene. The dominance of planktonic diatom species during this time, primarily *Cyclostephanos*

*dubius*, is indicative of deep-lake environments (*sensu* Gasse *et al.*, 1989; Barker *et al.*, 1994a; Tapia *et al.*, 2003) and not only supports the interpretation of Builth *et al.* (2008), but also suggests a degree of lake sensitivity to climate forcing. Moreover, the continuous record throughout the LGM demonstrates a level of lake permanency shared only by the deeper Lakes Bullenmerri and Purrumbete. This permanency, in combination with low measured salinity in the present drought, suggests that groundwater may be an important source of water to the lake.

The age of the Builth *et al.* (2008) record supports the assertion of Head *et al.* (1991) that the lava-flows on the surrounding landscape are nearer to 30,000 years old than the 5,000 years, as suggested by Ollier and Joyce (1964). Both the lake and the lava-flows have an intrinsic cultural value to the local indigenous community. Where the lava flows have interrupted local drainage lines, swamps have formed, and these have been the site of landscape management and sustainable aquaculture by the local Gunditjmarra people for more than 5,000 years (Builth *et al.*, 2008). The individual swamps were joined by “an elaborate system of stone fish weirs, channels and retaining walls ... [and], as lake and feeder stream levels rose and fell seasonally ... , the fish (principally eels) were caught as the water was diverted through traps” (Horton, 1994:576). The success of the aquaculture sustained an industry in smoked eels and eel oil which formed the basis of sedentary settlement (Builth, 2004).

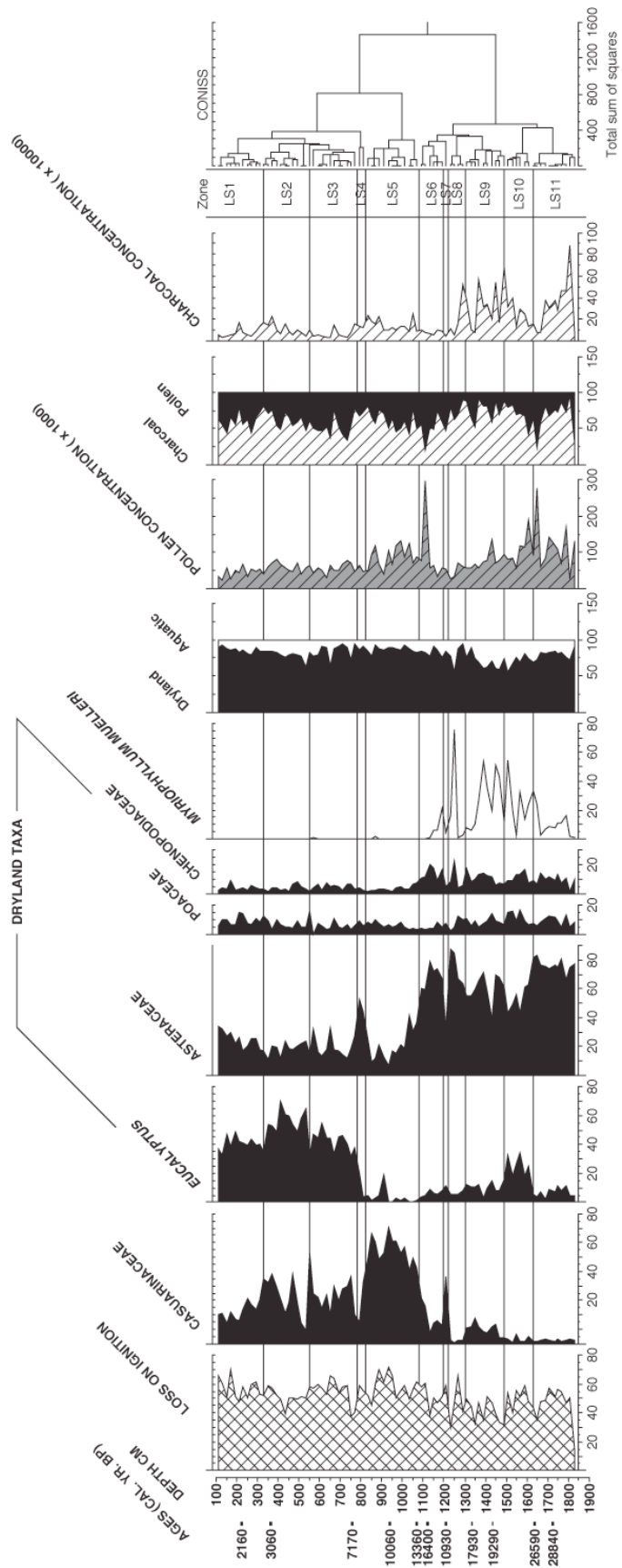


Figure 3.17: The Lake Surprise pollen record, from Builth et al. (2008)



NOTE:  
This figure is included on page 80 of the print copy of  
the thesis held in the University of Adelaide Library.

*Figure 3.18: The fine resolution pollen record from Lake Surprise core LSFS  
(Chris White, unpublished data)*

NOTE:

This figure is included on page 81 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 3.19: Relative abundance of main diatom species from the 13.5 m Lake Surprise core (Aline Philibert, unpublished data)*

NOTE:  
This figure is included on page 82 of the print copy of  
the thesis held in the University of Adelaide Library.

*Figure 3.20: Lake Surprise sedimentary mercury analysis from the combined  
18.5 m core and core LSFS (Left) and from core LSFS alone (Right). Note change  
of scale on horizontal axes. (Jacqui Hellings, unpublished data)*

## CHAPTER 4 – FIELD, LABORATORY AND STATISTICAL TECHNIQUES

This chapter is presented in three sections. The first (4.1) outlines field techniques including site selection for both the study sites and the modern calibration set, surface sediment retrieval and sediment coring methods. Section 4.2 focuses on laboratory techniques such as diatom sample preparation, species identification and enumeration, sediment analysis and sediment dating. The final section (4.3) is concerned with the statistical techniques used in the development and application of a diatom-conductivity transfer function.

### 4.1 Field techniques

#### 4.1.1 Selection of palaeoenvironmental study sites

The aim of this project was to examine, in fine resolution, the climate of the recent past in southern Australia using diatom-inferred salinity as the main indicator of changing moisture balance. In particular, the goal was to identify any evidence of the MWP and the LIA in Australia and, in the context of the current drought, to determine if regional moisture balance changed during these contrasting periods. It is known that in some locations, both the MWP and LIA have been associated with notable temperature (e.g., Holmgren *et al.*, 1999; Mann *et al.*, 2003, 2008; Moberg *et al.*, 2005) and hydrological (e.g., Fritz *et al.*, 1994; Laird *et al.*, 1996a; Nesje and Dahl, 2003) anomalies. Consequently, study sites were required which would reflect and record these potentialities.

Site selection is critical to achieving these aims and, as such, a number of criteria were applied to identify locations that would be most suitable to a high resolution reconstruction of palaeoclimatic change. These criteria are outlined below and summarised in Table 4.1.

In order to maximise the ability to reflect moisture balance changes, the first criterion was that the lakes are free from riverine input and have as small a catchment area as possible. This is to ensure that the lake is responding to localised factors, thereby removing the variables of landscape or regional scale factors which can confound interpretation of the record (Fritz, 2008). Ideally, the lakes should also be closed basins, free from interaction with groundwater. Lakes such as these have been described by De Deckker (1988) as being akin to palaeo-rain gauges due to their ability to record, in the

sedimentary archive, fluctuations in lake water salinity, a proxy for the precipitation/evaporation (P/E) ratio. This effectively narrowed the search to volcanic ‘crater lakes’ and, thus, to the volcanic plains of western Victoria, due to the number of such lakes in that region.

The second criterion was that the lakes are permanent water bodies. This was a requirement to ensure the production and maintenance of a continuous sedimentary and fossil record. In a fine-resolution study such as this, the loss of even a small part of the record due to sediment deflation from the lake drying out would greatly compromise any climatic reconstruction. Therefore, in the context of the current severe drought in the study area, and given the number of lakes in the western plains that had recently dried (see Table 4.2), it was assumed that, if a lake contained water under these current conditions it was, therefore, suitable for further consideration.

The third criterion, again in the context of the current drought, was that the ideal lakes would be within the oligosaline (0.5-5 g/L) to polysaline (20-30 g/L) (*sensu* Gasse *et al.*, 1987) range. The reason for this criterion is predominantly due to diatom ecology. As discussed in Section 2.5, diatom species that are present at higher salinities are generally tolerant of a broad range of salinities (e.g., Fritz, 1990). In developing a diatom transfer function from the lakes of the western plains of Victoria, Gell (1995) noted that the most significant changes in diatom species assemblages occurred between the oligosaline to polysaline ranges. Once lake salinity reached the eusaline range (30-40 g/L) and beyond, species diversity declined and the species that remained generally tolerated a much larger salinity gradient. In terms of their utility to palaeosalinity reconstructions then, these species, and therefore these sites, could not offer the same potential as those that preferred the fresher end of the scale.

The fourth criterion was that the lakes must have a relatively rapid sedimentation rate. Ideally, this would be a minimum of 1 metre per 1000 years in order to enable a sediment sampling regime which would provide, on average, a per sample temporal resolution of sub-decadal scale (assuming one sample per centimetre). A sub-decadal sampling regime was the minimum requirement in order to increase the potential of capturing any instability within the sedimentary record.

The final criterion was that the lakes must be climatically sensitive. That is, that the lakes should respond to, and reflect, climatic forcing. Several crater lakes in western Victoria have previously been demonstrated to meet this criterion, including Lakes Keilambete (Bowler, 1981; Chivas *et al.*, 1986; Mooney, 1997), Bullenmerri and Gnotuk (De Deckker, 1982). However, these lakes failed to meet one or more of the other criteria (see Table 4.1) and were eliminated from consideration. Wantzen *et al.* (2008:151) state that “The degree by which ... water-level fluctuations influence lake systems strongly depends on lake morphology, where shallow lakes or those with large shallow margins are the most sensitive.” Taking this into consideration, and after implementing the above criteria, six lakes remained as potential study sites; Lake Tooliorook, Lake Rosine, Lake Purrumbete, Lake Bullenmerri, Lake Elingamite and Lake Surprise (see Figure 3.8 for location).

Table 4.1: Lakes considered for inclusion in this project and reasons for exclusion. X – denotes that criterion was not met. ? – denotes ‘unknown’.

Lake	CRITERIA FOR INCLUSION						References
	‘Crater Lake’	Permanent	< 30 g/L Salinity	Sedimentation rate 1m/1000 yrs	Diatoms present	Known to be climatically sensitive	
Keilambete	✓	✓	✗				Jones <i>et al.</i> (2001)
Gnotuk	✓	✓	✗				Timms (1976)
East Basin	✓	✓	✗				Gell (1995)
West Basin	✓	✓	✗				Gell <i>et al.</i> (1994) Gell (1995)
Tower Hill	✓	?	✓	✗			D’Costa <i>et al.</i> , (1989) Barry <i>et al.</i> (2005)
Mumblin	✓	✓	✓	✗			Timms (1975)
Tooliorook	✓	✓	✓	?	✗		
Bullenmerri	✓	✓	✓	?	?	?	Jones <i>et al.</i> (2001)
Purrumbete	✓	✓	✓	?	✓	✓	Tibby & Tiller (2007)
Rosine	✓	✓	✓	?	?	?	
Elingamite	✓	✓	✓	✓	✓	✓	Maddox (1967), Hussainy (1969), Yezdani (1970), Tudor (1973), Timms (1977), Mann (1991)
Surprise	✓	✓	✓	✓	✓	✓	Timms (1975), Tibby <i>et al.</i> (2006), Builth <i>et al.</i> , 2008

Lake Tooliorook was excluded after examination of an exploratory 80 cm sediment core found no fossil diatoms. Lake Rosine was excluded due to the presence of a thick weed mat across the entire lake floor making retrieval of a sediment core problematic. In addition, the generally low landscape relief suggested that the lake could easily

overflow in humid periods, allowing a net loss of salts and thereby jeopardising the potential for salinity reconstructions. Lakes Bullenmerri and Purrumbete were excluded as Lake Bullenmerri is known to have overflowed in the historic past (Sutcliffe in Jones, *et al.*, 1993) and is fed by groundwater springs (Jones, *et al.*, 2001) while Lake Purrumbete overflows into the Curdies River (Quinn *et al.*, 1996). In addition, both sites are deep, without large shallow margins (Timms, 1976) and, in light of Wantzen *et al.*'s (2008) summary of lake sensitivity noted above, it was considered that these sites may not be sensitive to low-frequency climatic changes. Thus, Lake Elingamite and Lake Surprise remained as the two most suitable sites.

Comparisons between previous studies of Lake Elingamite demonstrated a decline in maximum lake depth from 12 m in 1966 (Hussainy, 1969) to 9.2 m in 1976 (Timms, 1977) (see Section 3.4.1) which, when considered in the context of the gradual increase in Victorian annual mean temperature since 1970 (see Figure 3.4), was suggestive of a climatic response. Water abstraction may have contributed to this decline (Mann, 1991), however lake modelling (Rodger Jones, unpublished data) also suggested that declines in lake depth were linked to climatic forces. There is an overflow point on the north-western side of the crater (Figure 3.10), however it was approximately 12 metres higher than the lake level at the time, which, when viewed in conjunction with the *ca.* 2 km diameter of the lake, suggested that a substantial increase in moisture availability would be required in order for the lake to overflow.

At Lake Surprise, previous investigations had demonstrated both a suitable sedimentation rate and an indication, over longer time periods at least, that the lake was responsive to climatic forcing (Aline Philibert, pers. comm.). Having met all the criteria outlined above, coring of the lake was undertaken (see Section 4.1.3.2).

Therefore, all the above criteria had been met, with the only exception being a known suitable sedimentation rate. The only previous work in which sedimentary studies had been undertaken on this Lake Elingamite was that of Yezdani (1970). However, in that study, a short core was retrieved from the lake edge and no dating was undertaken to determine a chronology. Therefore, a lake-centre sediment core was taken to ascertain an indication of sedimentation rate (see Section 4.1.3.1). Sediment from the base of the

resulting 1.8 m core returned a date of  $2032 \pm 61$   $^{14}\text{C}$  yr BP (see Section 5.1.1), thus indicating a sedimentation rate appropriate for this study. Sediment samples were taken at 1 cm intervals and prepared for diatom analysis (Section 4.2.1.2). A rapid enumeration of these preliminary diatom samples confirmed that diatoms were present throughout the sedimentary record and that there were substantial changes in the fossil diatom record. Sediment was also examined for the presence of fossil ostracods (De Deckker, 1982) which, in appropriate conditions, can provide an indication of past salinity and temperatures (Chivas *et al.*, 1986). It was hoped that this would provide further insight into the palaeoclimate of the region, though unfortunately, no ostracods were preserved in the sediment.

Thus, after following strict criteria to identify potential sites suitable to high resolution study, Lakes Elingamite and Surprise became the study sites chosen for diatom-inferred palaeosalinity reconstructions. Given that the set of criteria was designed to screen out less suitable sites, then Lakes Elingamite and Surprise remain the most suitable sites, within one of the most suitable regions in Australia, to derive high-resolution diatom-based reconstructions of past salinity, and so a detailed record of climate change and variability.

#### 4.1.2 Modern diatom dataset field methods

As noted in Section 1.3 a diatom-salinity transfer function has previously been developed from lakes in the study region by Gell (1997). While this dataset covers a broad range of salinity (0.5 g/L to 133.0 g/L; mean: 9.72 g/L), reconstructions derived by applying the transfer function to preliminary fossil diatom counts from Lake Elingamite, an oligosaline site (*sensu* Gasse *et al.*, 1987), had very poor modern analogues. This was primarily due to much of the fossil diatom record from Lake Elingamite being dominated by the planktonic species *Discostella stelligera* and *Discostella pseudostelligera*, neither of which are in the modern dataset of Gell (1997). One potential reason for the absence of these species is that modern samples for the Gell (1997) dataset were mainly taken from the littoral zone rather than lake centre, thereby preferentially excluding deeper water (planktonic) species such as the *Discostellas*.

It was therefore apparent that Gell's (1997) transfer function, despite being derived from the same study region, would not provide an accurate reconstruction of



palaeosalinity from deep, oligosaline lakes such as those identified as the logical study sites. Consequently, in order to better characterise fresher sites, increase the potential for good modern analogues and strengthen the validity of palaeosalinity reconstructions from the study sites, it was decided to develop a new dataset in which diatom samples would be taken from lake centre surface sediments. In addition, lake centre plankton samples would be also be taken, while lake-edge epipelagic samples would be included where possible. The criteria for sites to be included in this new dataset are outlined below.

#### *4.1.2.1 Modern dataset site selection*

With the knowledge that the study sites were currently fresh to oligosaline in nature (*sensu* Gasse *et al.*, 1987), and mindful of the fact that they may have had higher salinities in the past, the lakes to be sampled for the new ‘fresh dataset’ were selected primarily based on their known salinities (from Gell, 1995 and Gell, pers. comm.) and their geographical location within the western plains of Victoria. Lakes which fell above the mesosaline threshold (20 g/L) were excluded. Similarly, lakes with periodic or permanent inflow from rivers or streams were excluded to better capture the diatom flora of lakes, rather than rivers.

Having met the criteria outlined above, a total of 44 lakes were identified as potential sampling sites. At the time of sampling, in December 2004, the region had been in drought for nearly a decade and, of the 44 lakes identified, only 19 proved suitable (Table 4.2). The remaining 25 were excluded as they had become too saline ( $n = 9$ ), were inaccessible ( $n = 1$ ) or had completely dried out ( $n = 15$ ). Interestingly, all 15 of the dry lakes had contained water ten years earlier (Gell, 1997), providing stark evidence of the impact of the current drought.

The resulting 19 lakes would be insufficient for accurate statistical reconstructions and fell well short of the 30 sites suggested by Birks (1998) as the minimum required for a training set. Consequently, in an effort to increase the size of the dataset, a second batch of samples were sourced from the Arthur Rylah Institute for Environmental Research (ARI) in The Department of Sustainability and Environment, Victoria. These samples formed part of a large study by the ARI of 200 surface sediment samples from 58 lakes and wetlands, primarily within the Wimmera region of western Victoria (see Figure 3.1). Researchers took a number of samples from random locations from within each

lake, identified using random number tables (Sabine Schriber, pers. comm.). Each sample was accompanied by individual water chemistry measurements from the exact site of sampling. Due to the often wide variation of these measurements, each site of sampling was considered to be a unique sample for inclusion in the dataset. Of the initial 200 samples, 86 met the project criteria of ‘fresh’ sites (< 20 g/L salinity) that were on public land<sup>3</sup>. The names, locations and site codes of these are presented in Table 4.3.

In addition, Mr. Chris Gouimanis, a researcher at the Australian National University whose study site was Blue Lake, the deepest maar lake in the region, contributed a surface sediment sample that would represent the deep, fresh end of the environmental gradient. Water chemistry data for Blue Lake was provided by the EPA of South Australia, who regularly monitor the lake’s chemistry. With the addition of the Wimmera and the Blue Lake samples, the calibration set consisted of 106 sites.

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<sup>3</sup> The ARI permitted the use of this data on the proviso that no wetlands situated on private property would be included in the study (Sabine Schreiber, pers. comm.).

Table 4.2: The 44 lakes visited for the original dataset and reasons for exclusion. (\* - denotes that salinity was greater than 20 g/L)

<i>Lake</i>	<i>Sample Code</i>	<i>Latitude</i>	<i>Longitude</i>	<i>Included/ Excluded</i>	<i>Reason for exclusion</i>	<i>Salinity (g/L)</i>
<b>Bolac</b>	<b>9a</b>	37.44	142.52	<b>Included</b>		6.533
<b>Bullenmerri</b>	<b>15a</b>	38.15	143.05	<b>Included</b>		8.747
Burn		38.17	143.32	Excluded	Too saline*	
<b>Burrumbeet</b>	<b>10a</b>	37.29	143.38	<b>Included</b>		5.49
Carpolac		36.51	141.19	Excluded	Dry	
<b>Charlegrark</b>	<b>1a</b>	36.46	141.14	<b>Included</b>		0.441
Cobrico swamp		38.18	143.00	Excluded	No Access	
<b>Cogambul</b>	<b>6a</b>	36.57	141.24	<b>Included</b>		4.502
<b>Colac</b>	<b>12a</b>	38.18	143.36	<b>Included</b>		2.324
Coradgill		38.06	143.21	Excluded	Dry	
Coragulac		38.16	143.30	Excluded	Too saline*	
Deep		37.55	143.10	Excluded	Dry	
<b>Elingamite</b>	<b>17a</b>	38.21	143.00	<b>Included</b>		2.117
<b>Freshwater</b>	<b>8a</b>			<b>Included</b>		1.549
Glenthompson		37.34	142.33	Excluded	Dry	
Gnalinegurk		38.16	143.30	Excluded	Dry	
Gnarputt		38.03	143.23	Excluded	Dry	
<b>Immensal</b>	<b>3a</b>	36.59	141.21	<b>Included</b>		6.741
<b>Immensal North</b>	<b>4a</b>	36.59	141.21	<b>Included</b>		9.681
Jaracter		37.03	141.31	Excluded	Dry	
<b>Linlithgow</b>	<b>7a</b>	37.45	142.13	<b>Included</b>		8.522
Miga		36.56	141.37	Excluded	Dry	
Milangil		38.07	143.12	Excluded	Too saline*	
<b>Modewarre</b>	<b>11a</b>	38.15	144.06	<b>Included</b>		8.845
<b>Mumblin</b>	<b>18a</b>	36.19	142.54	<b>Included</b>		0.631
Natimuk		36.42	141.56	Excluded	Dry	
Ondit		38.15	143.36	Excluded	Dry	
Punpundal		38.07	143.22	Excluded	Too saline*	
Purdigulac		38.15	143.30	Excluded	Dry	
<b>Purrumbete</b>	<b>16a</b>	38.16	143.13	<b>Included</b>		0.49
Round		38.07	143.12	Excluded	Too saline*	
Saddlers		38.14	143.05	Excluded	Dry	
Streatham		37.41	143.00	Excluded	Dry	
<b>Struan</b>	<b>14a</b>	38.01	143.25	<b>Included</b>		0.459
<b>Surprise</b>	<b>19a</b>	38.03	141.55	<b>Included</b>		0.043
Swanee		37.01	141.43	Excluded	Dry	
Tatutong				Excluded	Too saline*	
Terang Goodwitch				Excluded	Too saline*	
Terangpom		38.08	143.19	Excluded	Dry	
<b>Tooliorook</b>	<b>13a</b>	37.58	143.16	<b>Included</b>		4.721
Tower Hill		38.19	142.21	Excluded	Too saline*	
<b>Wallace</b>	<b>2a</b>	37.01	141.17	<b>Included</b>		1.337
Weering		38.05	143.40	Excluded	Too saline*	
<b>Yallakar</b>	<b>5a</b>	36.56	141.23	<b>Included</b>		2.818

Table 4.3: Lake names (where known), sample codes and locations of surface sediment samples provided by the Arthur Rylah Institute

Lake (or location)	Sample Code	Latitude	Longitude	Lake (or location)	Sample Code	Latitude	Longitude
Unknown*	34A/1	-37.04	141.32	Lake Cogumbul Wildlife Reserve	153/1	-36.95	141.41
	34A/2	-37.04	141.32		153/2	-36.95	141.41
	34A/3	-37.04	141.32		153/3	-36.95	141.41
Toolondo State Forest	56B/1	-37.04	141.83		153/4	-36.95	141.41
	56B/2	-37.04	141.83	Konnepra State Forest	195/1	-36.94	141.59
Toolondo State Forest	51/1	-37.05	141.83		195/2	-36.94	141.59
	51/2	-37.05	141.83	Unknown	1000/1	-36.91	141.28
	51/3	-37.05	141.83		1000/2	-36.91	141.28
Toolondo State Forest	56/1	-37.04	141.83		1000/3	-36.91	141.28
	56/2	-37.04	141.84	McGlashin Swamp Wildlife Reserve	1002/1	-37.09	141.76
	56/3	-37.04	141.84		1002/2	-37.09	141.76
	56/4	-37.04	141.83		1002/3	-37.09	141.76
Unknown	93/1	-37.07	141.27		1002/4		
	93/2	-37.07	141.27	Unknown	2101/1	-37.06	141.29
Unknown	106/1	-37.00	141.59		2101/2	-37.06	141.29
	106/2	-37.00	141.58		2101/3	-37.06	141.29
	106/3	-37.00	141.58		2101/4	-37.06	141.29
	106/4	-37.00	141.58	Champion Swamp Wildlife Reserve	3094/1	-36.97	141.35
Lumeah Swamp Wildlife Reserve	107/1	-37.00	141.29		3094/2	-36.97	141.35
	107/2	-37.00	141.29		3094/3	-36.97	141.35
	107/3	-37.00	141.29		3094/4	-36.97	141.35
	107/4	-37.00	141.29	Parsons (Collins) Lake	7000/1	-37.00	141.35
	107/5				7000/2	-37.01	141.35
Toolongrook State Forest	115/1	-37.04	141.68		7000/3	-37.01	141.35
	115/2	-37.04	141.68		7000/4	-37.00	141.35
	115/3	-37.04	141.68	Lake Bringalbart	7011/1	-36.83	141.17
	115/4	-37.04	141.68		7011/2	-36.84	141.16
Kurrayah Swamp Wildlife Reserve	119/1	-37.08	141.29		7011/3	-36.84	141.16
	119/2	-37.08	141.29		7011/4	-36.84	141.16
	119/3	-37.08	141.29		7011/5	-36.83	141.16
Yallakar State Forest	136/1	-37.11	141.45	Alakilu Swamp Wildlife Reserve	9003/1	-36.92	141.38
	136/2	-37.11	141.45		9003/2	-36.92	141.38
Yallakar State Forest	136B/1	-37.11	141.45		9003/3	-36.93	141.38
	136B/2	-37.11	141.45		9003/4	-36.92	141.38
Wartook State Forest	137/1	-37.03	142.31	Lake Kemi Kemi Wildlife Reserve	9007/1	-37.07	141.29
	137/2	-37.03	142.31		9007/2	-37.07	141.30
	137/3	-37.03	142.30		9007/3	-37.07	141.29
	137/4	-37.03	142.31		9007/4	-37.07	141.30
Meereek Flora Reserve	139B/1	-37.09	141.18		9007/5	-37.07	141.30
	139B/2	-37.09	141.18		9007/6	-37.07	141.30
	139B/3	-37.09	141.18	Unknown	9053/1	-36.71	141.81
	139B/4	-37.09	141.18		9053/2		
Unknown	144/1	-37.07	141.47				
	144/2	-37.07	141.47				

#### 4.1.2.2 *Water sampling*

Collection of water samples for the ‘fresh dataset’ occurred at the same time, and from the same location within the lake, as the modern diatom surface sediment samples (Section 4.1.2.3). Göransson *et al.* (2004) found that lake-centre water samples can be considered to be more representative of surface water chemistry than ‘lake-edge’ samples and so, where possible, mid-lake sampling was undertaken. A minimum of 500 ml was collected from within the photic zone in a container which had previously been washed in a dilute hydrochloric acid solution. The container was capped while under water to exclude air bubbles. A second, smaller, water sample was collected at the same time and passed through a 45 µm filter into a clean collection tube. All samples were labelled and kept cool until they were transported to the Analytical Services Unit of the Commonwealth Scientific and Industrial Research Organisation (CSIRO) Land and Water division in Adelaide for detailed analysis using the standard methods of the American Public Health Association (1998).

Field measurements were also taken at the time of diatom sampling and water collection. These consisted of secchi depth, sampling depth, conductivity, dissolved oxygen, pH and water temperature. With the exception of secchi and sampling depth, these measurements were taken from within the photic zone with a TPS 90-FL Field Lab which was calibrated prior to each use. Secchi depth was measured as the depth at which a secchi disc was no longer visible from the lake surface. Sampling depth was measured with a weighted tape measure. As outlined above, water chemistry measurements for the Wimmera samples, and for Blue Lake, were provided by the relevant Government agencies.

#### 4.1.2.3 *Diatom sampling and surface sediment retrieval*

As mentioned above, the rationale for developing a new training set from fresh lakes was primarily due to poor analogues in the existing Gell (1997) training set for the preliminary fossil diatom species assemblages from core LE1 and the need for a greater representation of planktonic species. Therefore, surface sediment were taken from the deepest part of the lake. In the absence of detailed bathymetric measurements, samples were taken from the centre of the lake under the assumption that this location would have the least likelihood of benthic or littoral species and greatest likelihood of planktonic species. Samples from Lakes Bullenmerri and Purrumbete, with depths of approximately 60 metres and 45 metres respectively, were taken from nearer to the lake

edge at depths of approximately 5 metres (Lake Purrumbete) and 10 metres (Lake Bullenmerri) due to the limitations of the sampling equipment.

Surface sediments samples for the original dataset were collected using a modified Hongve surface sediment corer (Wright, 1990) with a clear Perspex coring barrel. The cores were brought to the surface and extruded to the point where the sediment/water interface was just below the top of the core barrel. Using a large bore 50 ml medical syringe, the sediment/water interface was drawn off and extruded into a sample pot. Approximately 10 ml of 100% ethanol was added to each sample to preserve the diatom species assemblage. Samples were kept cool during transport back to the laboratory.

#### 4.1.3 Sediment core retrieval

There are numerous methods of retrieving sediment cores for analysis. Which method is employed is determined by factors such as the type of sediment, the depth of the lake, the aims of the study, the biological proxy being examined and, often, by the resources available. In all cases however, the common goal is to retrieve the desired amount of sediment while maintaining the stratigraphic integrity of the sediment column. For reasons outlined below, each lake in this project required the use of a different coring method.

##### *4.1.3.1 Lake Elingamite*

Due to the depth of the lake (approximately three metres), and the available resources, a rod-less piston corer (Chambers and Cameron, 2001) was used to core Lake Elingamite in October, 2003. Bathymetric measurements of Timms (1977) demonstrated that the lake has a generally broad, funnel-shaped floor (see Figure 3.13). Using this as a guide, and taking regular depth measurements with a weighted tape measure, the deepest section of the lake was located as a coring site. Once the boat was stabilised using three anchors in a triangular formation, the corer was used to drive a 2 m length, 60 mm internal diameter, PVC pipe into the sediment. The leading edge of the pipe had been sharpened to ease entry into, and reduce the risk of disturbing, the sediment. The weights were removed and replaced with a retrieval mechanism to bring the core to the surface where the base was sealed. Once the coring head had been removed, the excess water was siphoned off and the pipe cut to a length just above the surface sediments.

The pipe was then capped, sealed and kept in an upright position for transportation back to the laboratory.

#### *4.1.3.2 Lake Surprise*

Two cores from Lake Surprise were used in this project. The cores were taken from the deepest part of the lake which was determined by using a weighted tape measure while using the bathymetric measurements of Timms (1975) as a guide (Figure 3.16). A large pontoon was placed over the coring location and secured in place by four ropes affixed to the lake shore.

The first core (LST1), measuring 231 cm, was taken by a rod-less piston corer (Chambers and Cameron, 2001) in the same manner as described for Lake Elingamite. In order to capture the entire sediment/water interface with intact stratigraphy, a second core (core LSFS measuring 110 cm) was taken using a 'frozen spade' corer (Neale and Walker, 1996). Frozen spade coring freezes the sediment to metal plates through the use of 'dry ice' (frozen CO<sub>2</sub>) and is particularly useful in locations such as Lake Surprise where there is a large sediment/water interface which can not be collected through other coring means (Neale and Walker, 1996).

## **4.2 Laboratory techniques**

The techniques employed in the laboratory consisted of the preparation of diatom slides from both modern and fossil samples, analysis of sediment, the identification and enumeration of diatom taxa, sieving for *Pinus* pollen grains (core LE1 only) and radiometric dating.

### 4.2.1 Diatom sample preparation

Several methods exist for preparing and fixing diatoms onto microscope slides for enumeration, such as Hasle and Fryxell (1970), Battarbee (1986), Renberg (1990) and Pientiz and Smol (1993). These differ primarily in the amount of original sample required the materials and chemicals used, and the length of time it takes to prepare the sample. In general though, all methods aim to isolate the diatom valves from extraneous materials within the sediment with minimal damage or chemical dissolution occurring to the valves as a result. While all the aforementioned methods achieve this, those most

commonly used, and those employed in this study, are the methods of Battarbee (1986) and Renberg (1990). Both employ the same general principle of H<sub>2</sub>O<sub>2</sub> and HCl digestion with the main differences between the two methods being the amount of original sample required and the laboratory equipment used.

#### *4.2.1.1 Modern dataset sample*

Due to the available sample volume, the method of Battarbee (1986) was employed. This method can be quicker than that of Renberg (1990) and is best employed where a greater amount of original material is available. Samples were placed into labelled glass beakers and placed in a hydrosonic bath for a maximum of 10 seconds. This serves two functions. The first is to disaggregate sediment, thus providing more surface area and therefore increasing the speed of chemical reactions when H<sub>2</sub>O<sub>2</sub> and HCl are added. The second is to break apart colonies of diatoms and thereby decrease the likelihood of large numbers of frustules lying in a 'girdle-view' orientation on the microscope slide, which can hinder identification. Following the method of Battarbee (1986), dilute hydrochloric acid (10%) was added to the beakers which were then placed onto a hotplate and warmed for three hours to remove carbonates. The samples were then left to settle overnight before the supernatant was carefully poured off. A series of three rinses with distilled water followed, with a minimum of six hours between each rinse and removal of the supernatant. After the final rinse, the samples were treated with 10% hydrogen peroxide and warmed for three hours to remove organic matter before repeating the settling and rinsing process. Distilled water was added to the final sample until the opacity was deemed to be adequate to ensure an appropriate density of valves. A pipette was used to place an aliquot of 800 µl onto a microscope coverslip which was then left to air dry overnight. Once dried, the coverslip was inverted onto a drop of Naphrax mounting medium on a warmed microscope slide. The slide was heated to remove toluene bubbles from the Naphrax before being placed on a cool surface where the coverslips were further pressed to remove all air bubbles and to place the diatoms into a valve-view orientation. This method of mounting samples onto microscope slides was employed for all diatom samples used in this project.

#### *4.2.1.2 Fossil diatom samples*

Due to the fine resolution nature of the study, a large number of fossil diatom samples were prepared. It was imperative that the amount of sediment used for these analyses be kept to a minimum to preserve the remaining sediment for further analyses such as bulk



*Pinus* sieving (Lake Elingamite only), radiocarbon dating and sedimentary analysis. For these reasons, the Renberg (1990) method was employed as it requires the use of smaller amounts of original material and large numbers of samples can be prepared simultaneously using the same chemicals as the method of Battarbee (1986). Furthermore, by preparing samples in 10 ml centrifuge tubes, accurate and uniform sample volumes could be derived, thus enabling quantitative diatom abundance measurements.

As the process of sediment sampling and preparing quantitative samples differed slightly between sites, each site is discussed separately below.

#### *4.2.1.2.1 Lake Elingamite core diatom sampling*

In order to remove the sediment from the coring pipe, an electric circular saw was used to cut the pipe longitudinally along two sides. A guide box was constructed in order to stabilise the saw, ensure a precise depth of cutting and avoid damage to the sediment. A broad flat blade was then pressed through the two cuts, bisecting the sediment which was then laid out onto a flat surface lined with a clean plastic sheet. A measuring tape was placed along each side of the core and affixed to the sheet to avoid movement (Figure 4.1). Using the measuring tapes to guide precise locations, a length of fishing line was stretched between the measuring tapes and the sediment was cut horizontally into 0.5 cm slices. A fine metal spatula was used to cut away the outermost layer of sediment which may be a source of contamination from younger sediments as the core was driven down into the sediment during the coring process. The slices of sediment which remained were placed in labelled sample pots. Both the spatula and the fishing line were thoroughly cleaned between samples and the remainder of the core which was not being sampled was covered with cling film to avoid desiccation.

Samples for diatom analysis were taken through the individual slices of sediment using a cut-off 1 ml medical syringe, thereby ensuring contiguous sampling along the entire core. For each slice approximately 0.2 cm<sup>3</sup> of sediment was taken and added to a pre-weighed, 10 ml centrifuge tube which was then re-weighed to determine the wet weight of the sediment sample being prepared.



*Figure 4.1: Core LE1 having been split lengthways and laid out for subsampling into 0.5cm slices. A fine bladed spatula is used to remove the outer layer of sediment, a possible source of contamination.*

Following Renberg (1990), approximately 3 ml of 30% hydrogen peroxide was added to the centrifuge tubes which were gently agitated on a daily basis until no further reaction (bubbling) was evident, suggesting that digestion was complete. The samples were then rinsed by removing the supernatant using gentle suction, taking care not to remove any of the sample, and re-filling the tubes with distilled water before gently agitating the tube to resuspend the sample. This process was repeated three times with samples left to stand undisturbed for a minimum of 24 hours between rinses. Following digestion by  $H_2O_2$  the samples were clear and it was decided the HCL treatment would not improve the sample quality, so this step was omitted.

After the final rinse, microscope slides were prepared by removing supernatant until approximately 3 ml remained in the tube. Using a vortex stirrer, samples were agitated before distilled water was added to bring the volume to 10 ml. From this, and while the sample was still in suspension, a 200 µl aliquot was removed using a pipette and added to a 22 mm x 22 mm microscope coverslip. A further 800 µl of distilled water was added to the coverslip to ensure that the sample would completely cover the entire surface area and, thereby, provide an equal distribution of diatom valves across the coverslip. These were then left to air dry before being fixed onto a microscope slide with Naphrax as described above.

#### *4.2.1.2.2 Lake Surprise core diatom sampling*

Two cores were examined from Lake Surprise; LSFS, a frozen spade core of the uppermost 110 cm of sediment and sediment/water interface, and a hammer driven piston core (LST1) of 231 cm in length.

Core LSFS was the subject of a collaborative study with researchers from Monash University in Melbourne. Once retrieved, the core was kept frozen on a bed of dry ice for transportation back to Monash University laboratories where it was stored at approximately -30°C. While still in a frozen state, the core was sliced into 1 cm slices using a Dremel hand-held MultiPro rotary tool with a diamond tipped blade. The resulting samples were stored in labelled pots and allowed to defrost in a refrigerator. The very high moisture content of the samples enabled the use of a medical syringe to aspirate 1 ml of the agitated sample for diatom sample preparation. This was added to a 10 ml centrifuge tube and the samples were prepared using the method of Renberg (1990), as outlined above. In this instance (and for core LST1) a water bath was used to heat samples for 2 hours at 85°C, thus significantly accelerating the preparation process. As above, the HCl step was omitted as the samples were clear following H<sub>2</sub>O<sub>2</sub> digestion. When preparing individual slides, a 40 µl aliquot (thereby equating to the density of the Lake Elingamite samples) was taken from the final sample volume of 10 ml and added to a coverslip containing 1 ml of distilled water. In the event that this did not provide a sufficient coverage of diatom valves for efficient enumeration, a second coverslip was prepared in the same manner with an 80 µl aliquot added. Dried coverslips were mounted onto slides as described above.

Core LST1 was split lengthways, laid out and sub-sampled into 1 cm slices in the same manner as described for core LE1 above. As with Lake Elingamite, samples were taken using a cut-off medical syringe which was pushed through the sediment slices to provide a contiguous sampling regime along the core. The samples were individually weighed and were prepared as above. The HCl step was included for this core though, due to elevated sedimentary carbonate content in parts of the core.

#### 4.2.2 Diatom identification and enumeration

Identification of diatoms was undertaken on a variety of microscopes at either 1000x or 1500x magnification. For consistency, the same microscope was used for each core. For example, all the samples from cores LE1 and LSFS were counted at 1500x magnification on a Nikon Eclipse E600 with Nomarski differential interference contrast, while all modern samples were counted at 1000x magnification on a Zeiss Axiolab using phase contrast. All diatom identification was undertaken at a minimum of species level.

A number of taxonomic texts were used to aid in identification of species. Primarily these were Krammer and Lange-Bertalot (1986, 1988, 1991a, 1991b), Sonneman *et al.* (1999), Lange-Bertalot (1996, 1998) and occasionally Foged (1978). In 1993, Droop *et al.* stated that “published floras ... are few, often outdated, inconsistent, incomplete or present other problems, for example linguistic”. While there has been an improvement in the quantity and quality of texts since that time, the nomenclature for diatoms remains less than clear. Different systems of nomenclature have evolved through divergence of opinion, and which system is adopted varies within and between continents, countries, regions and even laboratories.

Generally speaking, there are two main systems of nomenclature adopted, one based on the works of Krammer and Lange-Bertalot (1986, 1988, 1991a, 1991b) and the other on the proposed modifications of genera by Round *et al.* (1990) and numerous further modifications. In this project, the most recent names and authorities have been used. Taxonomic authorities and synonyms of the most common fossil species identified are presented in Appendix 1.

#### 4.2.2.1 Modern samples

In order to determine the number of valves which provided the transfer function with the greatest accuracy and smallest errors, modern samples were counted to cumulative totals of 100, 200, 300 and 400 valves. Counting was undertaken along coverslip transects and, to accelerate the counting process, samples which failed a criterion of a minimum of 10 valves per transect were considered barren and not included in the dataset. Data analysis showed that a count of 200 valves was sufficient to provide a strong model with little model improvement arising from counting more valves (see Section 5.4.1).

#### 4.2.2.2 Fossil samples

Increased concentrations of fossil diatoms, as a proxy for palaeoproductivity, have been associated with periods of warmer palaeoclimates in certain circumstances (e.g., Smith, 2002; Heroy *et al.*, 2008). Therefore, from the outset, it was decided to quantitatively measure diatom concentration. Approximately five to ten fields of view were enumerated along a transect before moving to the next transect and thereby eliminating the risk of counting the same field of view twice. All diatoms which lay entirely within the field of view were enumerated and the distance between both the fields of view and the transects, was random. The number of fields of view (FoVs) required to attain the total valve count (see below) was used to determine diatom valve density using the formula of Smith (2002);

$$\left( \left( \frac{\text{Size of coverslip}}{\text{Size of FoV}} \right) \times \left( \frac{\text{No. of valves counted}}{\text{No. of FoVs}} \right) \times \left( \frac{\text{Volume of slurry}}{\text{Volume of aliquot}} \right) \right) \div \text{Dry weight of sample} = \text{Valves per dry gram of sedimen}$$

Battarbee (1986) recommends a minimum of 300 valves are counted to obtain a fully representative sample of diatom valves in the sample. However, following on from the results of analysis of the modern samples outlined above, and given the low species diversity evident in preliminary counting of core LE1, it was hypothesised that a representative sample could be attained, and the counting process thereby accelerated, by counting a lower number of valves. This is based on the assumption that the number of species identified is inversely proportionate to the number of valves counted. Moreover, any new species identified would be of little value when applied in the transfer function. To test this hypothesis, counts of 100, 200, 300 and 400 valves were

undertaken at five regularly spaced intervals along the length of the core and the results applied to the following formula of Pappas and Stoermer (1996), where efficiency is defined as “the probability that a new species encountered is minimal”.

$$\text{Counting efficiency} = 1 - \frac{\text{Number of species}}{\text{Number of individuals}}$$

Results from core LE1 indicated that, while there was a notable increase in counting efficiency between counts of 100 valves and 200 valves, there was little benefit to be gained by increasing counts from 200 valves to 300 or 400 valves (Figure 4.2). Indeed, the lowest efficiency value after 200 valves were counted was 0.92 from a depth of 20 cm. The finding that counting 200 valves can provide a representative sample is consistent with the findings of Bate and Newell (1998) who applied the same system to diatom samples from Kiewa River in Victoria. However, Bate and Newell (1998) caution that a count size of 200 valves may not be adequate in all situations, and samples with a more diverse diatom community, or numerous species occurring in similar numbers, may require a greater count size to provide a representative sample. With this in mind, the same method was applied to cores LSFS (Figure 4.3) and LST1 (Figure 4.4). In both cases, results, once again, indicated that a total count of 200 valves would provide a sample with suitably high counting efficiency.

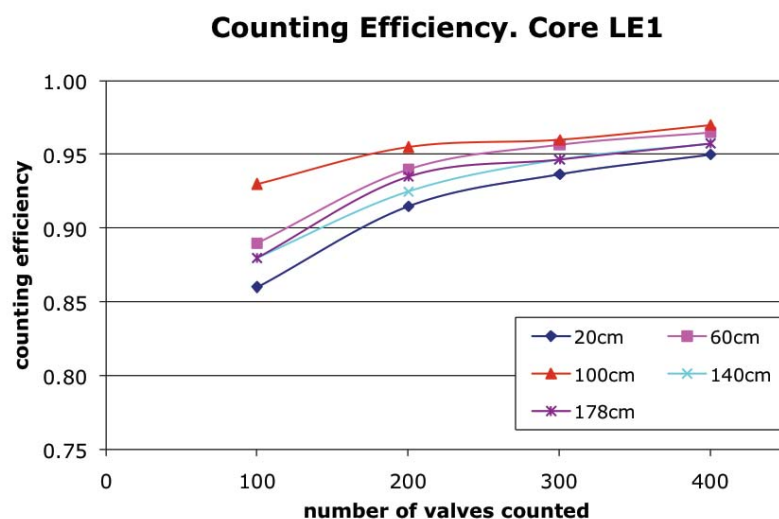


Figure 4.2: Determination of counting efficiency from core LE1

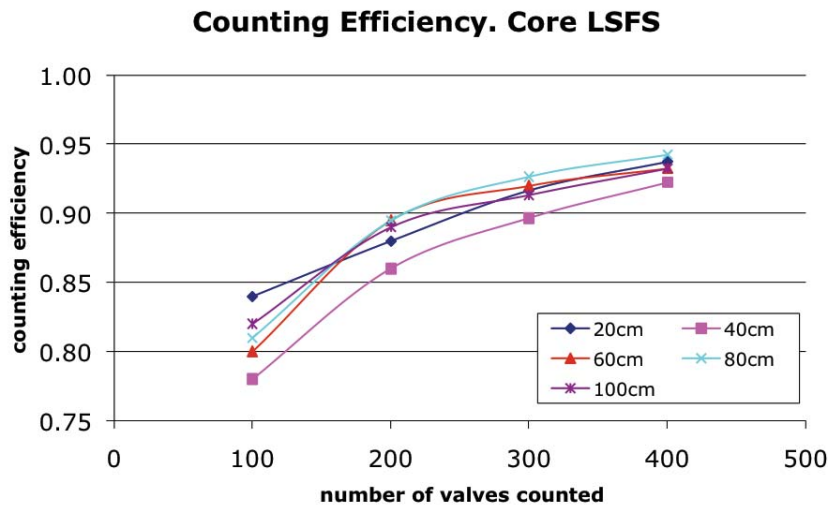


Figure 4.3: Determination of counting efficiency from core LSFS

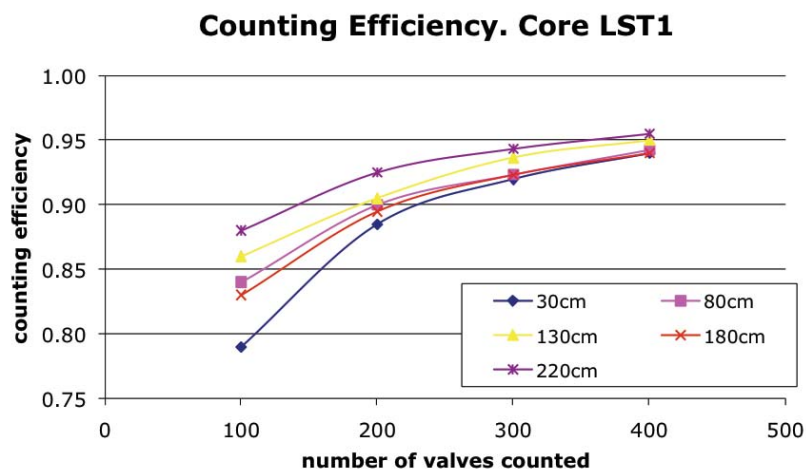


Figure 4.4: Determination of counting efficiency from core LST1

Raw species counts were entered in to Microsoft Excel (v. 12.0) before being converted to proportions and transferred to the program C2 (v. 1.5) (Juggins, 2007). Core zonation was determined using the Constrained Incremental Sum of Squares (CONISS) function in TGView (v. 2.0.2) (Grimm, 1991-2001), using Euclidean distance as a measure of similarity and being stratigraphically constrained. Stratigraphic diagrams were developed using the programs TGView (v. 2.0.2) (Grimm, 1991-2001) and C2 (v. 1.5) (Juggins, 2007), with final illustrations prepared in Adobe Illustrator (v. 13.0). Fossil diatom counts, as relative abundance data, for cores LE1, LSFS and LST1 are presented in Appendices 2, 3 and 4 respectively.

#### 4.2.3 Sediment analysis for exotic *Pinus* pollen

*Pinus* is an introduced tree genus in Australia and, as such, is commonly used in Australian palynological studies as an indication of the arrival of, or alteration of the landscape by, Europeans (e.g., Dodson *et al.*, 1993; Kershaw *et al.*, 2004b; Bickford and Gell, 2005). While it is often the case that the appearance of *Pinus* in the sedimentary record occurs some time after the initial vegetation change brought about by Europeans (Gell *et al.*, 1993; Mooney *et al.*, 2000), it can still provide supplementary and supportive information when used in conjunction with radiometric dating methods. In the case of this study, it was intended as an independent tracer to supplement and potentially validate  $^{210}\text{Pb}$  dating (Smith, 2001). This investigation was only implemented on sediment from Lake Elingamite as detailed pollen analyses was undertaken on the Lake Surprise LSFS core as part of the collaboration with researchers at Monash University, Melbourne, while core LST1 was assumed to commence before the period of European arrival.

The method of bulk sieving for *Pinus* follows that of Ogden (1996). This method is designed as a rapid means of isolating *Pinus* pollen from sediment without the need for the more detailed chemical pre-treatment methods usually employed for pollen analysis (e.g., Moore *et al.*, 1991). All sampling was undertaken in a positively pressurised room to reduce the risk of contamination from airborne pollen. A sample of 1 cm<sup>3</sup> of sediment was taken from core LE1 at depths 20 cm, 30 cm, 40 cm, 50 cm and 60 cm using a medical syringe with the tip removed. The samples were individually dispersed in 60-100 ml of hot 10% potassium hydroxide (KOH) and gently stirred with the use of a magnetic flea before being left to soak overnight. The next day, the dispersed sediments were passed through a 65 µm sieve and placed in a large, one litre beaker. The fraction retained in the sieve was thoroughly rinsed, with the runoff collected in the beaker. The contents of the beaker were then passed through a 38 µm sieve and the contents of the sieve retained and placed into a graduated sample tube. An aliquot of 200 µl of the sample was measured, using a graduated pipette, placed onto a microscope slide, covered with a microscope coverslip and examined with a Zeiss Axiolab light microscope at 200x magnification for the presence of *Pinus* grains. The aim was purely to determine presence or absence of *Pinus* in the record, so quantitative analysis was not undertaken. A blank sample, which was run concurrently with the sediment samples to check for possible contamination, was devoid of *Pinus* grains.



Commencing with a sample from 20 cm, and moving down core, the apparent density of *Pinus* grains gradually diminished until no *Pinus* grains were identified in the sample from a depth of 60 cm. At this point, further sediment samples were taken at 1 cm intervals from a depth of 51 cm to 59 cm and the process was repeated. Again, the apparent density of *Pinus* grains gradually diminished down core until they became very rare in sample 55 cm. At this stage, the entire 1 ml of sample was examined, one 200 µl aliquot at a time, finding only 4 grains. The process was repeated for samples from 56-59 cm with samples at 57 cm, 58 cm and 59 cm all being devoid of *Pinus* pollen. In this manner then, the depth of 56 cm was identified as the first arrival of *Pinus* pollen in the sedimentary record of Lake Elingamite.

#### 4.2.4 Sediment analysis

Sediment from cores LST1 and LSFS were examined at contiguous 1 cm levels for bulk density, moisture content, organic content and carbonate content. Samples from core LE1 could not be examined due to the fact that, despite the 0.5 cm slices of sediment being sealed in sample pots and refrigerated, the samples had desiccated before analyses could be undertaken.

Samples for LST1 were taken using a cut off medical syringe and 2 cm<sup>3</sup> of sediment was taken where sufficient sediment permitted. Samples from 1 cm to 15 cm on core LSFS consisted of only 1 cm<sup>3</sup> of sediment due to the smaller size of the original sediment sample. No measurements were taken on core LSFS at depths of 2 cm, 61 cm, 105 cm and 110 cm, due to insufficient material.

Bulk density was determined by dividing the weight of the sample by the mass of the sample. Moisture content, expressed as %H<sub>2</sub>O, was determined by calculating the percentage of wet weight lost after the sample had been in the oven for a minimum of 12 hours at 105°C. Organic content was estimated after burning the sediment in a muffle furnace at 550°C for two hours and calculated using the following formula of Heiri *et al.* (2001);

$$\text{LOI}_{550} [\% \text{organic}] = ((\text{DW}_{105} - \text{DW}_{550}) / \text{DW}_{105}) \times 100$$

where  $DW_{105}$  represents the dry weight, in grams, of the sample before combustion and  $DW_{550}$  represents dry weight, in grams, of the sample after combustion. Similarly, the carbonate content of the sediment, expressed as % Inorganic Carbonate, was determined after burning the sediment for 2 hours at 950°C and calculated using the formula of Heiri *et al.* (2001);

$$LOI_{950} [\% \text{inorganic carbonate}] = (((DW_{550} - DW_{950})/DW_{105}) \times 100) \times 1.36$$

where the factor of 1.36 is included to account for the difference in molecular weight between carbonate and carbon dioxide (Heiri *et al.*, 2001).

In terms of calculating carbonate content, Heiri *et al.* (2001) found that burning the sediment at 950°C for 2 hours was “sufficient to evolve all the carbon dioxide from the carbonate”. In terms of calculating organic content, Boyle (2004) suggests that a higher temperature of 550°C, as opposed to the temperature of 375°C proposed by Beaudoin (2003) for example, is preferred in situations where the sediment is organic-rich and clay-poor, as is the case in Lake Surprise. Yet factors such as exposure time, sample size, the position of crucibles in the furnace (Heiri *et al.*, 2001), and the presence of clays and salts (Santisteban *et al.*, 2004), have been found to influence results. Any resulting errors would, therefore, be compounded when calculating carbonate content as  $DW_{550}$  is a requirement of the formula. Estimating the amount of organic content through LOI is, therefore, “essentially a compromise” (Boyle, 2004:125) and is best viewed as a qualitative, rather than quantitative, data (Santisteban *et al.*, 2004). Raw data was entered into Microsoft Excel (V.10) with graphs generated using C2 (V. 1.5) (Juggins, 2007), with final illustrations prepared in Adobe Illustrator (V. 13.0).

#### 4.2.5 Dating of sediments

As the focus of the project was fine-resolution climatic reconstruction, the aim was to support this with a fine-resolution chronology, thereby determining such phenomena as the duration of wet/dry phases, while also allowing correlation with, and comparison to, other well dated records from the region (e.g., Cook *et al.*, 2000). This required dating methods for both the longer term, around 2000 years, and the short-term, approximately 150 years, which would encompass the period of European settlement in the region (Jones, 1999). Older samples were dated using either conventional radiocarbon dating

or Accelerated Mass Spectrometry (AMS), while younger sediments were dated primarily using radioisotopes  $^{210}\text{Pb}$  and  $^{137}\text{Cs}$ . All dating for this project was undertaken at either Australian Nuclear Science and Technology Organisation (ANSTO) laboratories (Fink *et al.*, 2004) in Sydney, Australia (sample labels commencing with 'OZ'), or at the Waikato Radiocarbon Dating Laboratory at the University of Waikato, New Zealand (sample labels commencing with 'Wk').

#### 4.2.5.1 Lake Elingamite

Once core LE1 had been retrieved, a sample of sediment was sent for dating to determine whether the core was likely to be suitable for fine-resolution studies. Standard radiocarbon dating was used in this instance, as opposed to AMS, due to financial constraints. The results (Wk-12384) confirmed a suitable age and so further samples were submitted for dating. The location of these samples was guided by notable bio-stratigraphic changes identified from preliminary diatom counts. Initially, with the diatom data showing a major and unprecedented change around the depth of 55 cm, this level was estimated to be boundary of European impact (approximately 150 years) and was therefore not sampled for radiocarbon dating due to the large fluctuations in the calibration curve through this period (McCormac *et al.*, 2004; Reimer *et al.*, 2004) and the resulting difficulties in calibrating such young radiocarbon dates.

Accelerator mass spectrometry (AMS) dating was employed for these samples and the reasons for this are two-fold. Firstly, AMS dating requires smaller sample size than conventional radiocarbon dating, therefore minimising the destruction of the core and maximising the ability to meticulously locate the samples along the core. Secondly, in an attempt to increase the precision of the dates and reduce, or eradicate, in-lake processes from affecting results, it was decided to date the pollen fraction of the sediment and, due to the small sample sizes, AMS dating was required. Brown *et al.* (1989) were the first to use pollen grains for dating purposes using AMS and state that "dates obtained by this method should provide more reliable radiocarbon chronologies for paleo-environmental studies than have been obtainable by bulk sediment dating". The method of abstracting pollen concentration from the sediment follows that of Killian *et al.* (2002).

As mentioned above, younger sediments were dated using  $^{210}\text{Pb}$ . In the first instance, ten samples were taken at 5 cm intervals between the depths of 0 and 45 cm and

submitted for the gamma-spectrometry assay. This concurrently measures  $^{137}\text{Cs}$ , a radioisotope associated with atmospheric testing of nuclear devices in the 1950s. The limit of  $^{210}\text{Pb}$  dating is roughly 150 years depending on the interpretive model employed (Binford, 1990) and, given the knowledge that *Pinus* first appeared in the record at 56 cm, it was estimated that a sample from a depth of 45 cm would approximate the chronological limits of the dating procedure. However, results from the upper samples were difficult to explain and results from the lower samples surprisingly suggested that the first appearance of  $^{137}\text{Cs}$  associated with the 1950s atmospheric testing had not yet been captured. In light of this, a further batch of eight samples, taken at 10 cm intervals from 0 – 70 cm, were submitted with the dual aims of testing the validity of the results from the first batch and determining the sedimentary level of European arrival. Due to a lack of remaining sediment though, sample sizes were insufficient for gamma spectrometry. Essentially, this meant that either the fine-resolution nature of the dating was compromised by ‘bulking up’ the samples with more sediment from above, or below, the sample, or the alpha-spectrometry assay was used, which precludes the measurement of  $^{137}\text{Cs}$  but can be undertaken on smaller samples. The decision was made to maintain the integrity of the fine-resolution dating and to use the alpha-spectrometry, at the expense of having  $^{137}\text{Cs}$  data, for the second batch of samples.

While the results of the second batch of  $^{210}\text{Pb}$  dating were more plausible than those of the first, they did not correspond to either the arrival of *Pinus* in the record, nor the chronology suggested by the  $^{14}\text{C}$ -AMS dates (see Table 5.6). A potential explanation for this discrepancy came in the dating report which stated that “very little to no pollen” was in the samples dated. In effect, the substance that had been dated was the organic matrix that remained after the sample had been treated for pollen analysis. Killian *et al.* (2002) found that a lack of purity in the pollen concentration, and admixtures from other organic remains, resulted in AMS dates approximately “660 years ‘too old’” (Killian *et al.*, 2002:26). In that instance, annual laminations of the lake sediments aided in determining the size of the error. Unfortunately, no such laminations exist in Lake Elingamite sediments.

In an attempt to determine if such an error did exist in this instance, a further three samples were sent to a different laboratory. One of these consisted purely of dried sediment and was dated using conventional radiocarbon dating. This sample (Wk-17870) was taken from 147-149 cm in order to contrast against the dated sample of

'organic-matrix' from 146 cm (OZH117). The aim was to determine if there was a difference between dating the organic fraction of the sediment and dating the organic fraction of the sediment which remained after the sediment had been prepared for pollen analysis. The other two samples were taken from 56-58 cm, a level chosen due to the proximity of the arrival of exotic *Pinus* pollen in the record (56 cm), which, it was hoped, could act as a potential independent tracer. The aim was to quantify any age difference between a 'pure' pollen sample and an 'organic matrix' sample. After both samples were treated for pollen analysis, as per the method of Killian *et al.* (2002), one sample (Wk-17822), consisting of the post-preparation matrix, was dated via AMS. The other sample (Wk-17821) was treated to further concentrate pollen through the use of a series of polytungstate densities. The sample was treated with progressively 'lighter' liquids, starting at 2.0 g/cm<sup>3</sup> and continuing down to as low as 1.1 g/cm<sup>3</sup>. The effect of this is to separate out extraneous material by floating off the lighter pollen. It is assumed that this method ensured the sample was as near to 'pure pollen' as is possible, a fact confirmed by the laboratory technician (Fiona Petchey pers. comm.). Confirmation was also received that the sample contained a substantial amount of pollen, suggesting laboratory error may have been responsible for the lack of pollen in the previous samples.

#### 4.2.5.2 Lake Surprise

Dating the sediments from Lake Surprise was straightforward in comparison to the Lake Elingamite record. Sediments from the top of core LSFS were dated using alpha-spectrometry due to the small sample masses of the frozen sediment. Samples were taken from depths of 3 cm, 5 cm, and 10 cm and, thereafter, at 10 cm intervals to 90 cm. This dating was undertaken at ANSTO by a researcher from Monash University as core LSFS was the subject of a collaborative study. Samples were prepared as per unpublished ANSTO methods ENV-I-044-001 (Alpha spectrometry), ENV-I-044-006 (Bulk iron removal by ether extraction), ENV-I-044-023 (Polonium analysis), ENV-I-044-027 (<sup>226</sup>Ra analysis) and ENV-I-044-031 (Sedimentation rate determination). Sediment ages were calculated using both the modified constant initial concentration (CIC) method of Brugam (1978), and the constant rate of supply (CRS) method of Appleby and Oldfield (1978).

Two samples of older sediments from LSFS were prepared for dating by the author, as were five samples from core LST1 (Table 4.3). The location of the samples was guided

by noted stratigraphical changes in the sediment. All samples were prepared by the author as per Killian *et al.* (2002) and checked microscopically for the presence of pollen grains prior to AMS being performed. Samples from both cores were taken from matching depths to aid in core correlation as it was assumed that the cores would overlap and could be correlated using a combination of radiocarbon dates, sediment analysis, core stratigraphy and diatom biostratigraphy. For reasons unknown (Quan Hua pers. comm.), the sample from the depth of 127-128 cm on core LST1 (OZI018) failed to return a reading. The replacement sample (OZI591) was taken from 80-81 cm. All Lake Surprise samples dated by AMS used the method of Hua *et al.* (2004) for the graphitisation stage as this has been shown to improve the reaction rate and efficiency when dating small samples.

Core	Sample Depth	Laboratory Code
LSFS	80-81cm	OZI015
LSFS	104-105cm	OZI016
LST1	80-81cm	OZI591
LST1	104-105cm	OZI017
LST1	127-128cm	OZI018*
LST1	141-142cm	OZI019
LST1	167-168cm	OZI020
LST1	214-215cm	OZI021

Table 4.4: Samples for  $^{14}\text{C}$ -AMS dating from Lake Surprise. (\*- indicates dating was unsuccessful on this sample).

### 4.3 Statistical techniques

The development of the diatom-based salinity transfer function was an iterative process in that the direction of further investigations was often determined by the results of previous analysis. Therefore, many of the detailed methods employed in the analysis and development of the transfer function are detailed in Chapter 5. The following description will therefore focus on the underlying reasons behind the employment of the analyses performed and the processes involved.

#### 4.3.1 Data screening

As outlined in Section 4.1.2.1 this project developed two datasets. The first, named 'WP' (denoting Western Plains), consisted of 20 sites (the original 19 sites, with the addition of a single sample from Blue Lake) that had full ionic chemistry measurements. The second dataset, containing 86 samples, emanated from the Wimmera region of western Victoria with samples provided by the Arthur Rylah Institute. However, the suite of environmental variables measured from the Wimmera sites differed from those measured for the WP sites. Of those variables absent from the Wimmera sites were measurements of ionic chemistry. The concentration and composition of major ions may be important in explaining diatom distribution (Gasse *et al.*, 1995; Gell, 1997; Davies *et al.*, 2002) and so the omissions of these measures limited the scope of analyses.

Once merged, the WP and Wimmera datasets had nine variables in common. When the two datasets were amalgamated a number of samples ( $n = 47$ ) in the combined dataset had missing values in one or more environmental variables. Missing values in ecological data matrices are not uncommon (Birks *et al.*, 1990; Legendre and Legendre, 1998) and there are a number of ways in which this problem can be addressed. Historical data can be used where it is known to be dependable (e.g., Gell, 1997) or the mean (e.g., Wunsam and Schmidt, 1995) or median values (e.g., Ryves *et al.*, 2002) of the variable can be substituted. Values can also be estimated through linear regression with significantly correlated variables (e.g., Reid, 1997; Tibby, 2000) though, as with the use of mean and median values, this may only be appropriate where a small proportion of sites are missing the variable.

In the combined dataset, a number of sites ( $n = 38$ ) had three or more of the nine measured variables missing. Replacement of missing variables was deemed inappropriate as too many variables were missing and the samples would effectively represent an average of the dataset. Moreover, due to the number of samples in question, the ability to cross-validate the model using leave-one-out methods would be compromised. Therefore, samples with three or more missing variables were deleted from the dataset prior to data exploration.

Several samples remained that had missing data for pH, temperature or depth. Missing pH ( $n=3$ ), temperature ( $n=8$ ) and depth ( $n=3$ ) data were substituted by the mean of that

variable across the entire dataset. In the case of depth, data for Blue Lake were removed prior to calculating the mean of the dataset as it was a significant depth outlier which skewed the data. As the C2 computer program, which was used to develop the transfer function, requires numeric data only (Juggins, 2007), measurements below detection limits were assigned a value of half that limit (e.g., measurements of  $< 0.1$  were assigned the value of 0.05).

An initial Detrended Correspondence Analysis (not shown), performed as discussed in Section 4.3.2, identified sample 9003/1, from Alakilu Swamp Wildlife Reserve, as a significant outlier that was located more than seven standard deviation units away from the next closest sample. This was due to the diatom assemblage of this sample being dominated (91%) by a single species (*Navicula salinicola*) that was only present in one other sample in the dataset (9003/2) at much lower relative abundance (11%). Sample 9003/1, as well as 'barren' samples in which no diatoms were identified, were removed from the dataset. Following this data screening, the final dataset, named 'Full', consisted of a total of 169 species from 47 sites with nine measured environmental variables.

Frequency distributions of the measured environmental variables showed that six variables (alkalinity, calcium, conductivity, depth, magnesium and phosphorus) were skewed. In order to lessen the influence of extreme measurements, these were all log-transformed, with the exception of calcium and phosphorus, which were transformed using a log+1 equation due to low initial measurements returning negative values under a standard log-transformation.

#### 4.3.2 Detrended Correspondence Analysis

A Detrended Correspondence Analysis (DCA) was performed on the 'Full' dataset using CANOCO for Windows version 4.5 (ter Braak and Šmilauer, 2002) to determine the length of the species gradient and also to gain a measurement of the total inertia, which provides an indication of the breadth of variability within the dataset. The results guide the next step of analysis. Linear responses to variables by the species, as indicated when the gradient length is less than 2 standard deviations (Birks, 1998), requires the use of linear indirect gradient techniques. A gradient length more than 3 standard deviations (ter Braak and Prentice, 1988) indicates that direct gradient analysis



techniques that assume a unimodal species response, such as Canonical Correspondence Analysis, are more appropriate. Results of DCA, detailed in Section 5.4.4, indicated that canonical correspondence analysis was warranted for this dataset.

#### 4.3.3 Canonical Correspondence Analysis

Having determined that techniques assuming a unimodal species response were most suitable, Canonical Correspondence Analysis (CCA) of the sites, species and environmental data was performed, using Hill's scaling, to examine species-environment relationships and identify the dominant environmental variable exerting the most influence over diatom species distribution. To assess the significance of the dominant variable, 9999 random Monte Carlo permutation tests were run. All analyses were undertaken using the program CANOCO for Windows version 4.5 (ter Braak and Šmilauer, 2002).

#### 4.3.4 Weighted Averaging regression and calibration

As discussed in Chapter 2, by using regression and calibration through the methods of Weighted Averaging – Partial Least Squares (WA-PLS) (ter Braak and Juggins, 1993), and its precursor, Weighted Averaging (WA), it is possible to reconstruct past environmental conditions from biostratigraphic records. The application of these methods to diatom species abundances is well established (Birks, 1998) and has been used to reconstruct various water quality variables including, among others, salinity (e.g., Gell, 1997; Taukulis and John, 2009) and conductivity (Ng and King, 1999; Ryves *et al.*, 2002; Tibby and Reid, 2004). All WA and WA-PLS analyses in this project were undertaken using the program C2 version 1.5 (Juggins, 2007).

The correlation coefficient and/or predictive error of transfer function models can be improved by removing the 'noise' of rare taxa and the extreme influence of outlying samples (Birks, 1998). Definitions of 'rare' taxa vary between studies and criteria for species deletion is often determined in an *ad hoc* manner (Wilson *et al.*, 1996). To determine the effect of various species deletion criteria, and identify the model with the best reconstructive capacity, five species datasets were developed, each with a different criteria for deleting species (Section 5.4.6). In the resulting models, outlying samples, defined as samples that had a jack-knifed residual value  $> \pm 1$  standard deviation of the

environmental variable being modelled (Gasse *et al.*, 1995; Jones and Juggins, 1995), were deleted iteratively before re-running the model. This continued until the improvement of the model, as measured by a decrease in the root mean squared error of prediction (RMSEP), was less than 5%. This value follows the ‘minimum adequate model’ approach advocated by Birks (1998).

#### 4.3.5 Modern Analogue Technique

To assess the potential robustness of the diatom-inferred reconstructions, the modern analogue technique was employed to measure the degree of similarity between the fossil samples and those in the calibration set. Using the ‘Modern Analogue Technique’ function in C2 version 1.5 (Juggins, 2007), squared chord distance was measured between each sample in the calibration set. The distance was then measured again between each fossil sample and the most similar sample in the modern dataset. Fossil samples with a minimum dissimilarity coefficient greater than, or equal to, the highest 10th percentile of the calibration set (Laird *et al.*, 1998; Kauppila and Valpola, 2003; Tibby *et al.*, 2003) were considered to have good analogues in the modern data set.

## CHAPTER 5 – RESULTS

This chapter presents the results of investigations of the Lake Elingamite and Lake Surprise cores (Sections 5.1 and 5.2 respectively), before addressing the correlation of Lake Surprise cores LSFS and LST1 in Section 5.3. Thereafter, Section 5.4 outlines the results of the development of a diatom-conductivity transfer function, while Section 5.5 presents the results of the application of the transfer function to fossil data in order to reconstruct past changes in conductivity. Finally, Section 5.6 evaluates the resulting reconstructions from the three cores by use of the modern analogue technique.

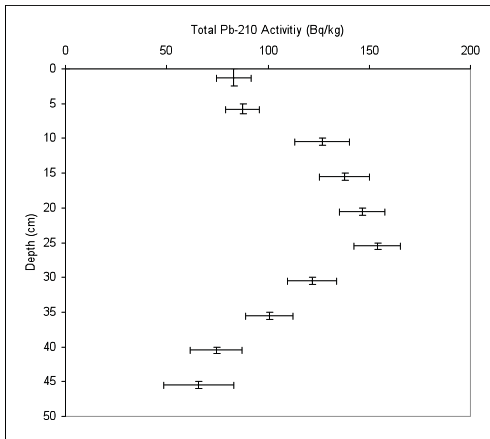
### 5.1 Lake Elingamite

#### *5.1.1 Chronology*

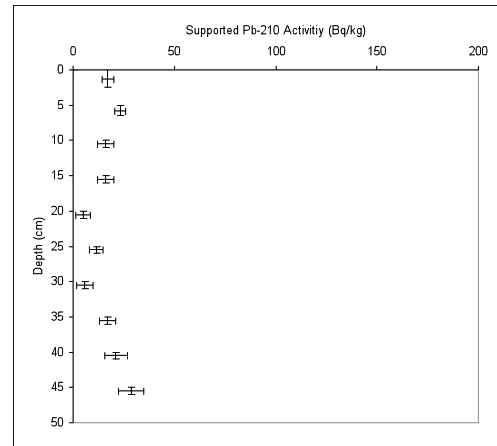
As outlined in Section 4.2.5.1, determining the chronology for core LE1 was not straightforward. Two attempts were made at determining an accurate chronology of the younger sediments at the top of the core. The first attempt employed gamma spectrometry, measuring  $^{210}\text{Pb}$  and using  $^{137}\text{Cs}$  as an independent marker. The results of this are presented in Table 5.1 with Figures 5.1 to 5.3 providing the graphical representation of the results.

*Table 5.1: Core LE1 Gamma spectrometry results*

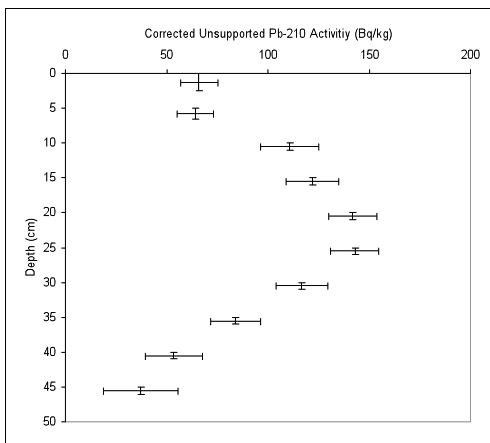
ANSTO ID	Depth (cm)	Count Date	Total Pb-210	Supported Pb-210	Unsupported $^{210}\text{Pb}$		$^{137}\text{Cs}$	
			(mBq/g) or (Bq/kg)	(mBq/g) or (Bq/kg)	Corrected to reference date 01-May-05 (mBq/g) or (Bq/kg)		Corrected to reference date 01-May-05 (mBq/g) or (Bq/kg)	
H731	0 - 3	12-May-05	83 ± 9	17 ± 3	66	± 9	10	± 3
H674	5 - 7	05-May-05	88 ± 8	23 ± 3	64	± 9	10	± 3
H732	10 - 11	13-May-05	127 ± 14	16 ± 4	111	± 14	9	± 4
H733	15 - 16	19-May-05	138 ± 12	16 ± 4	122	± 13	13	± 4
H734	20 - 21	26-May-05	147 ± 11	5 ± 3	142	± 12	9	± 3
H735	25 - 26	02-Jun-05	154 ± 11	11 ± 3	143	± 12	10	± 3
H736	30 - 31	08-Jun-05	122 ± 12	6 ± 4	117	± 13	14	± 4
H737	35 - 36	09-Jun-05	101 ± 12	17 ± 4	84	± 12	20	± 4
H738	40 - 41	14-Jun-05	74 ± 13	21 ± 6	53	± 14	14	± 4
H739	45 - 46	16-Jun-05	66 ± 17	29 ± 6	37	± 19	18	± 6



*Figure 5.1: Core LE1 total  $^{210}\text{Pb}$  ( $^{210}\text{Po}$ ) activity versus depth, gamma spectrometry results*



*Figure 5.2: Core LE1 supported  $^{210}\text{Pb}$  ( $^{226}\text{Ra}$ ) activity versus depth, gamma spectrometry results*



*Figure 5.3: Core LE1 total  $^{210}\text{Pb}$  ( $^{210}\text{Po}$ ) activity versus depth, gamma spectrometry results*

The results were interpreted as being indicative of sediment mixing in the top 25 cm (Atun Zawadzki, pers. comm.). Only the samples from 25 cm to 46 cm demonstrated a decreasing profile with depth and the sedimentation rate calculated from these samples ( $0.45 \pm 0.03$  cm/yr) was extrapolated to the top of the core. The resulting calculation of ages is presented in Table 5.2.

Table 5.2: Core LE1 gamma spectrometry sediment ages

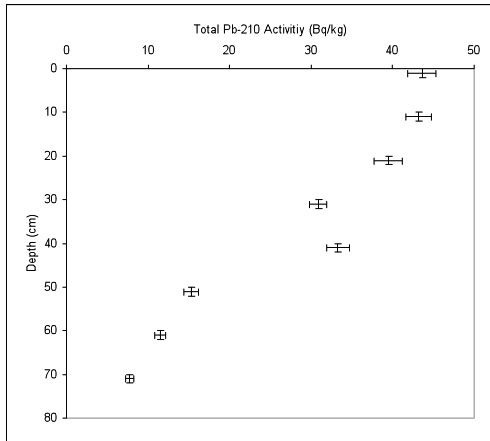
ANSTO ID	Depth (cm)	Combined Calculated CIC Ages (years)
H731	0 – 3	2.8 ± 28
H674	5 – 7	12.8 ± 1.9
H732	10 – 11	23.4 ± 5.0
H733	15 – 16	34.5 ± 2.8
H734	20 – 21	45.7 ± 3.5
H735	25 – 26	56.8 ± 4.3
H736	30 – 31	67.9 ± 5.1
H737	35 – 36	79.1 ± 5.9
H738	40 – 41	90.2 ± 6.7
H739	45 – 46	101.3 ± 7.5

Levels of measured  $^{137}\text{Cs}$  did not indicate a clear signal of the atomic testing period of the 1960s suggesting the  $^{137}\text{Cs}$  may be mobile in the sediment material and could therefore not be used to validate the interpreted  $^{210}\text{Pb}$  chronology (Atun Zawadzki, pers. comm.).

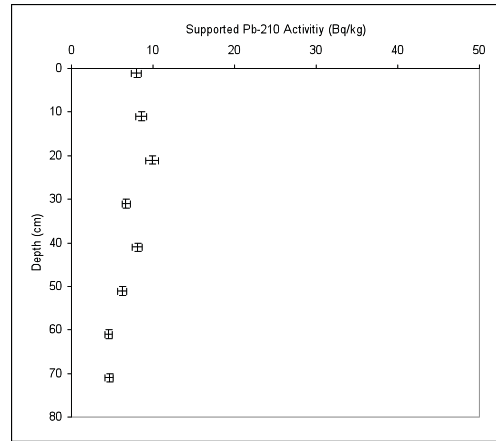
The chronological limit of  $^{210}\text{Pb}$  as a dating tool extends to roughly 150 years (Appley and Oldfield, 1978) which, in the context of western Victoria, is ideal for providing an indication of European arrival in the region which intensified after 1840 AD (Jones, 1999). In an attempt to capture this time frame, further samples were taken from the same core, though, in this instance, they were analysed by alpha spectrometry due to a smaller sample size requirement. The results, presented in Table 5.3 and illustrated in Figures 5.4 to 5.6, are notably different to those obtained from gamma spectrometry presented above.

Table 5.3: Core LE1 alpha spectrometry results and calculated ages.

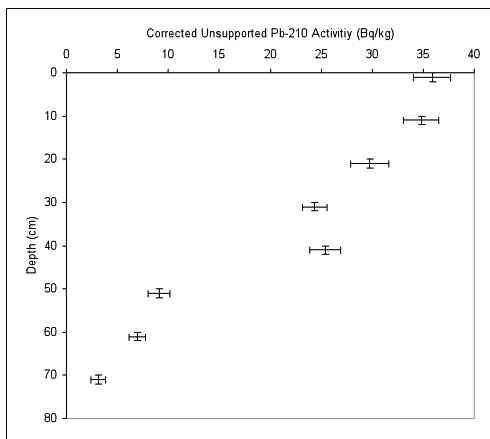
ANSTO ID	Depth (cm)	Total Pb-210	Supported Pb-210	Unsupported $^{210}\text{Pb}$	Combined Calculated CIC Ages (years)
		(mBq/g) or (Bq/kg)	(mBq/g) or (Bq/kg)	Corrected to reference date 01-May-05 (mBq/g) or (Bq/kg)	
I081	0 - 2	43.6 ± 1.7	8.0 ± 0.6	35.9 ± 1.8	0.6 ± 0.6
I082	10 - 12	43.1 ± 1.6	8.6 ± 0.6	34.8 ± 1.7	6.3 ± 0.7
I083	20 - 22	39.5 ± 1.7	9.9 ± 0.8	29.8 ± 1.9	12.1 ± 1.1
I084	30 - 32	30.9 ± 1.1	6.7 ± 0.5	24.3 ± 1.2	17.8 ± 1.5
I085	40 - 42	33.3 ± 1.3	8.1 ± 0.6	25.4 ± 1.5	23.5 ± 1.9
I086	50 - 52	15.3 ± 0.9	6.2 ± 0.5	9.1 ± 1.0	44.6 ± 3.6
I087	60 - 62	11.5 ± 0.7	4.6 ± 0.4	6.9 ± 0.8	65.6 ± 6.5
I088	70 - 72	7.7 ± 0.5	4.6 ± 0.5	3.1 ± 0.7	86.7 ± 9.6



*Figure 5.4: Core LE1 Total  $^{210}\text{Pb}$  ( $^{210}\text{Po}$ ) activity versus depth, alpha spectrometry results*



*Figure 5.5: Core LE1 Supported  $^{210}\text{Pb}$  ( $^{226}\text{Ra}$ ) activity versus depth, alpha spectrometry results*



*Figure 5.6: Core LE1 Total  $^{210}\text{Pb}$  ( $^{210}\text{Po}$ ) activity versus depth, alpha spectrometry results*

Largely due to the outlier at 42 cm, the results from alpha spectrometry indicated two zones of varying sedimentation rates, presented in Table 5.4.

Table 5.4: Calculated sedimentation rates for core LE1 based on alpha spectrometry results

Depth (cm)	Sedimentation rate (cm/yr)	Correlation coefficient ( $r^2$ )
0 - 42	$1.7 \pm 0.1$	0.9944
42 - 72	$0.5 \pm 0.1$	0.9575

As detailed in Section 4.2.5.1, several radiocarbon dates were obtained from Core LE1 with a variety of substances being dated. Results are presented in Table 5.5. Radiocarbon ages were calibrated using the Southern Hemisphere Calibration set (SHCal04) of McCormac *et al.* (2004) in the  $^{14}\text{C}$  calibration program Calib (version 5.0.1) and are presented as calibrated years before present (cal yr B.P.) and calendar years (AD/BC) (Table 5.6). Final ages were calculated by taking the midpoint between the two-sigma age ranges that had the greatest relative area under probability distribution. Calibration plots for all LE1 samples are presented in Appendix 5.

Table 5.5: Results of  $^{14}\text{C}$  analysis of samples from core LE1. All samples dated using AMS except WK-12384. Note: \* - denotes concentrated pollen sample as opposed to other samples labelled "Pollen" which had no pollen present after pre-treatment. \*\* - denotes the organic matrix which remained following pre-treatment for pollen analysis.

LAB. CODE	Sample Depth	Material dated	$\delta(^{13}\text{C})$ per mil	% Modern Carbon		Conventional $^{14}\text{C}$ age	
				pMC	1 $\sigma$ error	yrs BP	$\pm$
Wk-17821	56-58	Pollen*	$-23.5 \pm 0.2$	$86.3 \pm 0.4$		1181	37
Wk-17822	56-58	Organic**	$-22.2 \pm 0.2$	$92.5 \pm 0.4$		624	32
OZH245	76	"Pollen"	-21.0	86.88	0.49	1130	50
OZH244	82	"Pollen"	-25.00	88.99	0.51	940	50
OZH114	90	"Pollen"	-21.4	85.67	0.40	1240	40
OZH115	116	"Pollen"	-22.3	84.00	0.42	1400	40
OZH116	136	"Pollen"	-20.2	83.08	0.35	1490	40
OZH117	146	"Pollen"	-21.9	78.11	0.33	1980	40
Wk-17870	147-149	Organic	$-22.5 \pm 0.2$	$78.9 \pm 0.3$		1908	31
OZH118	170	"Pollen"	-19.6	79.90	0.37	1800	40
WK-12384	170-180	Organic	$-22.8 \pm 0.2$	$77.7 \pm 0.6$		2032	61

Table 5.6: Results of radiocarbon age calibration of samples taken from core LE1.

Sample	<sup>14</sup> C Age	One sigma age ranges	Relative Area under probability distribution	Two sigma age ranges	Relative Area under probability distribution	AGE cal yr B.P.	AGE AD/BC
<b>Wk-17821</b>	1181±37	975–1060	1	960-1140 1160-1168	0.985 0.015	1050	900
<b>Wk-17822</b>	624±32	545-563 602-628	0.433 0.566	529-572 589-645	0.439 0.560	617	1333
<b>OZH245</b>	1130±50	937-942 955-989 1031-1051	0.059 0.641 0.298	932-1013 1026-1055	0.738 0.262	972	978
<b>OZH244</b>	940±50	764-804 872-900	0.641 0.358	736-824 827-843 863-905	0.662 0.032 0.306	780	1170
<b>OZH114</b>	1240±40	1064-1096 1103-1138 1161-1168	0.459 0.159 0.081	1059-1172	1	1115	835
<b>OZH115</b>	1400±40	1275-1295	1	1187-1201 1257-1304	0.038 0.962	1280	670
<b>OZH116</b>	1490±40	1307-1339	1	1298-1360	1	1329	621
<b>OZH117</b>	1980±40	1828-1848 1862-1894	0.324 0.676	1822-1903 1907-1925	0.911 0.089	1862	88
<b>Wk-17870</b>	1908±31	1728-1824	1	1710-1873	1	1791	159
<b>OZH118</b>	1800±40	1617-1675 1686-1699	0.810 0.190	1570-1584 1539-1712	0.039 0.961	1652	298
<b>WK-12384</b>	2032±61	1834-1841 1865-1997	0.029 0.971	1742-1754 1781-1793 1811-2116	0.007 0.007 0.984	1963	-13

### 5.1.2 Fossil diatom analysis

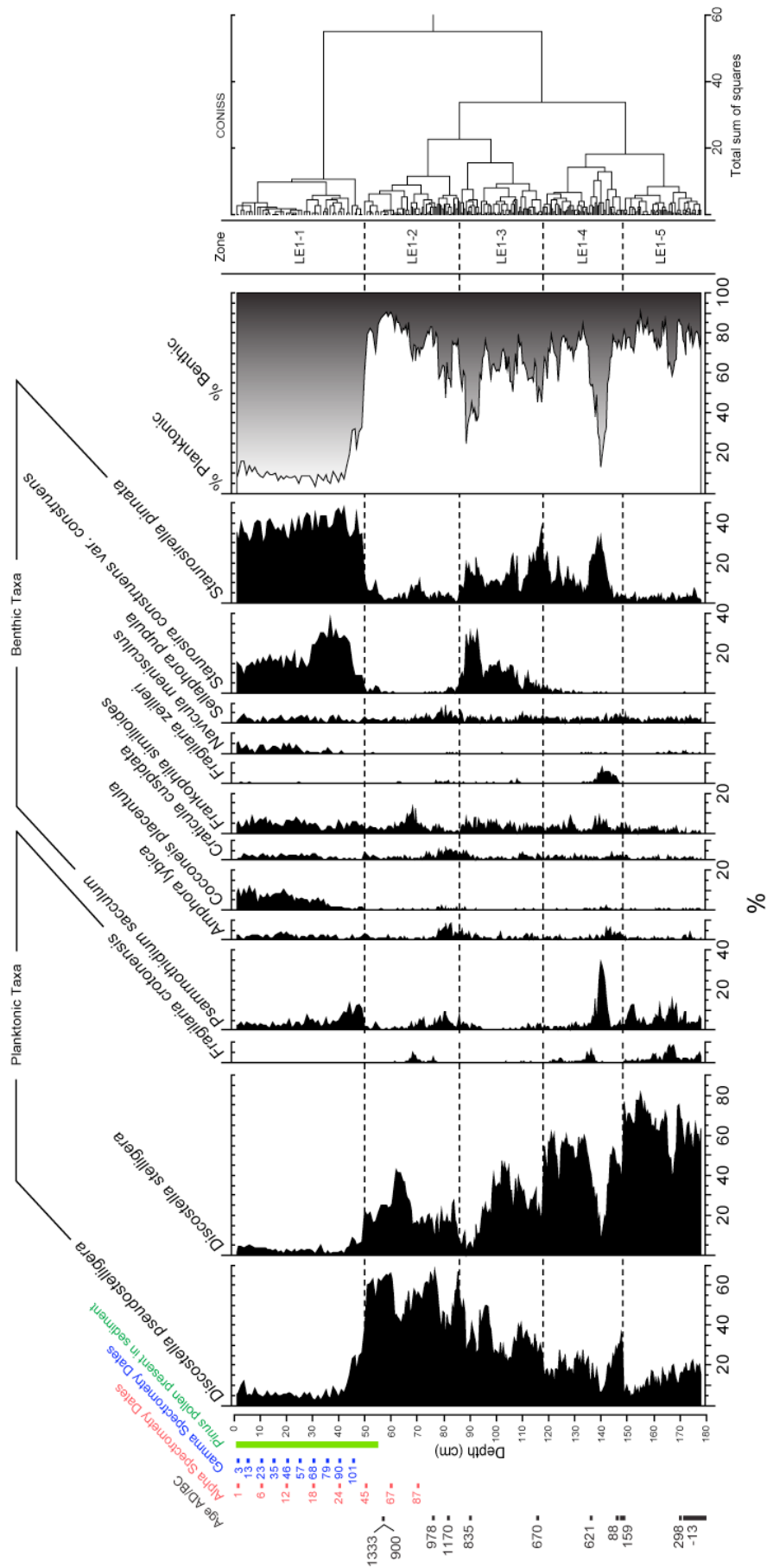
The results of the analysis of fossil diatoms from Core LE1 are illustrated in Figure 5.7. This figure represents all species with a representation of 5% or more in one sample. Using CONISS, five zones have been indentified which will be discussed below in reverse chronological order.

#### *Zone LE1-5 (178 – 148 cm)*

The zone is dominated by the planktonic *Discostella* genus. The gradual decline of *Discostella pseudostelligera* through the zone is mirrored by the gradual increase in *Discostella stelligera*. The benthic species of *Amphora libyca*, *Craticula cuspidata*, *Frankophilia similoides*, *Sellaphora pupula* and *Staurosirella pinnata* all maintain relatively low and constant numbers. *Fragilaria crotonensis* peaks to its highest representation in the record of approximately 8% from 166.5 cm to 168 cm which coincides with one of the peaks of *Psammothidium sacculum* which shows mild fluctuations across the zone.



Figure 5.7: Core LE1 Diatom stratigraphy showing all species with a relative abundance of 5% or more.



#### *Zone LST1-4 (148 cm – 118 cm)*

A rapid increase in proportions of *D. pseudostelligera* from 4.8% at 48.5 cm to 26.8% at 48 cm marks the beginning of the zone and is mirrored by an equally sharp decline in *D. stelligera*. A fall in the representation of both planktonic species follows and occurs in concert with a slight increase in proportions of *A. libyca* and *Fragilaria zeilleri*. These fluctuations precede the dominant feature of the zone, a sudden drop in the prominence of planktonic species primarily due to a rapid increase in the representation of *P. sacculum* and *S. pinnata*. Following this peak, the representation of *P. sacculum* gradually diminishes, though the relative abundance of *S. pinnata* continue to increase for the remainder of the zone. The proportion of the planktonic species also increase as *D. stelligera* resumes its dominance.

#### *Zone LST1-3 (118 cm – 86 cm)*

The increasing numbers of *S. pinnata*, seen in the previous zone, reach a peak at the base of this zone and account for 37.5% of the count at 117.5 cm. The proportions gradually decrease above this point though remain relatively constant with occasional fluctuations. A central feature of the zone is the steady increase in *Staurosira construens* var. *construens* which accounts for approximately a quarter of all valves counted from 93.5 cm to 88.5 cm. There is also a steady rise in the presence of *D. pseudostelligera* across the zone, while the representation of *D. stelligera* gradually decreases after peaking midway through the zone. Counts of *A. libyca*, *C. cuspidata*, *Frankophila similioides* and *Sellaphora pupula* all remain relatively constant at low levels.

#### *Zone LST1-2 (86 cm – 50 cm)*

Having been prominent in the previous zone, both *Staurosirella pinnata* and *Staurosira construens* var. *construens* are much less so in this zone, with the latter not being noted at all in the majority of samples. The zone is dominated by planktonic species with the main contribution coming from *D. pseudostelligera*. Small increases in *P. sacculum*, *A. libyca*, *C. cuspidata* and *Sellaphora pupula* precede the peak in *D. pseudostelligera*. As numbers gradually decline from this peak, small increases *Fragilaria crotonensis*, *Frankophila similioides* and *Staurosirella pinnata* precede a peak in *D. stelligera* which then gradually declines towards the top of the zone being replaced once again by *D. pseudostelligera*.

### *Zone LST1-1 (50 cm – 1 cm)*

The previous domination of the record by planktonic taxa rapidly diminishes at the beginning of the zone, though *D. pseudostelligera* declines at a slower rate, and maintains a higher representation for the remainder of the record, than *D. stelligera*. A major shift in the proportions of *S. pinnata* occurs at 49 cm increasing from 12.5% to 34.5% of the total. This is initially coincident with a small increase in the representation of *P. sacculum*, but this is soon replaced by a sudden rise in *Staurosira construens* var. *construens*. *Cocconeis placentula* and *Navicula menisculus* obtain their highest representation of the record in the upper half of the zone, coinciding with a reduction in the previously high numbers of *S. construens* var. *construens*. *Amphora libyca*, *Craticula cuspidata* and *Sellaphora pupula* which all maintain a low, relative constant, presence.

## **5.2 Lake Surprise**

Two cores were taken from Lake Surprise, core LSFS being a 'Frozen Spade' core of the top 110 cm and core LST1 being a longer 231 cm core of the deeper sediment. The results of investigations will be presented separately for each core.

### 5.2.1 Core LSFS

#### *5.2.1.1 Sediment composition*

The sediment lithology showed homogenous dark brown lake mud along the length of the core with no significant changes evident. This core was designed to capture the uppermost sediment and the sediment/water interface. As can be seen below, this resulted in a very high water content for the top samples and, once the sample was dried, a very small sample on which to undertake further analysis. This results in a high potential for error as the smallest change in the weight of the sample after treatment causes large fluctuations in the final calculations. Results of sedimentary analysis from the very top of this core should, therefore, be viewed with caution. At depths of 2 cm, 61 cm, 105 cm and 110 cm there was insufficient sediment to undertake analysis after diatom samples were taken.

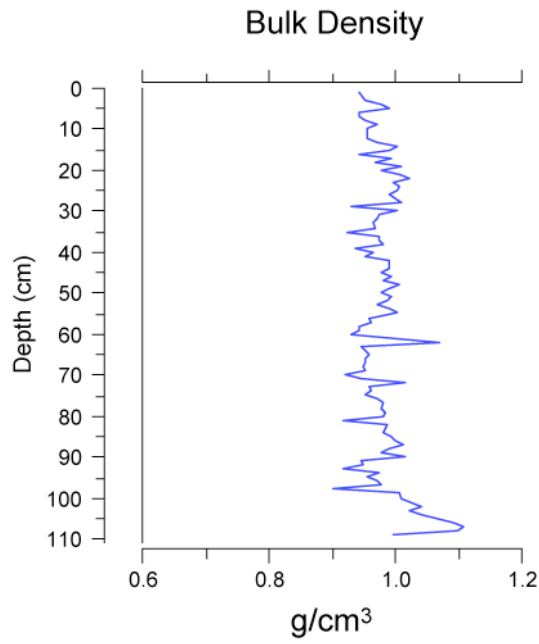


Figure 5.8: Bulk density calculations for Core LSFS

Bulk density fluctuates between measurements of 0.9 g/cm<sup>3</sup> at 98 cm and 1.1 g/cm<sup>3</sup> at 107 cm (Figure 5.8). The data are relatively constant, although is slightly elevated towards the base of the core. There is a slight increase in variability in the bottom 50 cm of the core.

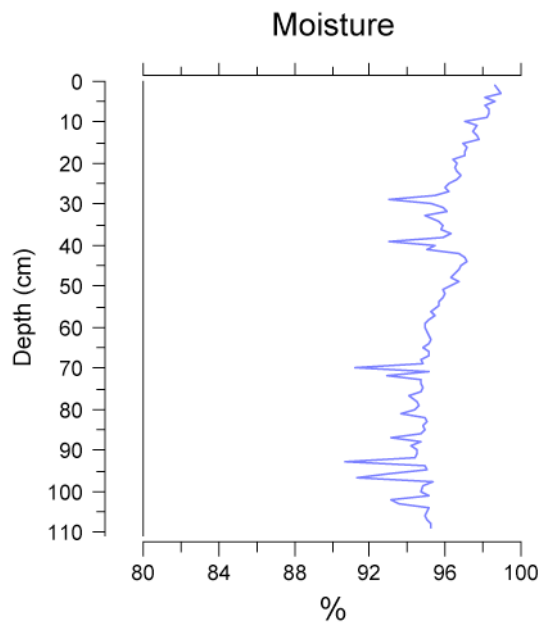
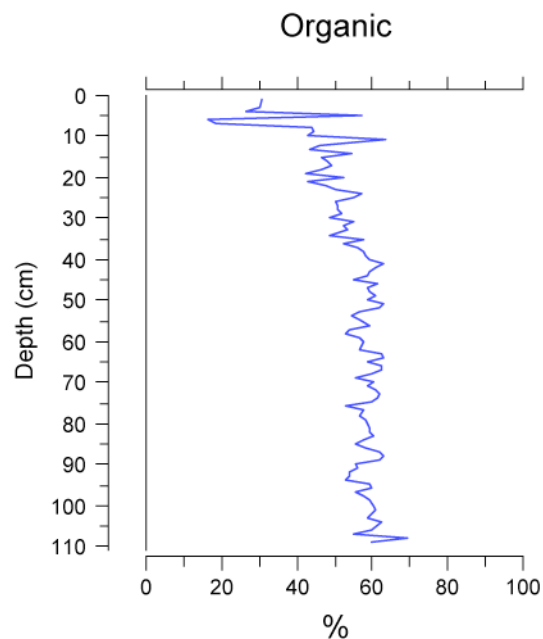


Figure 5.9: Moisture content calculations for core LSFS

The moisture content is very high in core LSFS (Figure 5.9). This is probably due to the frozen spade method of coring employed which freezes sediment and water to the plates of the corer, including that from the water column during ascent and descent of the corer. The general pattern over the bottom 50 cm of the core is stable, though variability is notably greater than in the upper 60 cm of the core, in which there is a gradually increasing moisture content. The upper 30 cm has a very high moisture content, attaining a maximum of 98.3% at 3 cm, indicative of the successful capture of the sediment/water interface.

From 40 cm to the base of the core the organic content is reasonably stable at around 60% with only minor variability (Figure 5.10). Notable fluctuations in the organic content of the sediment in the uppermost 20 – 30 cm of the core are probably unreliable due to the small sample size after drying.



*Figure 5.10: Sediment organic content for core LSFS*

As explained above, due to the small sample size, the results for inorganic carbonate content from the uppermost sediments should be viewed with caution. The remainder of the core shows some variability, though no obvious trends (Figure 5.11).

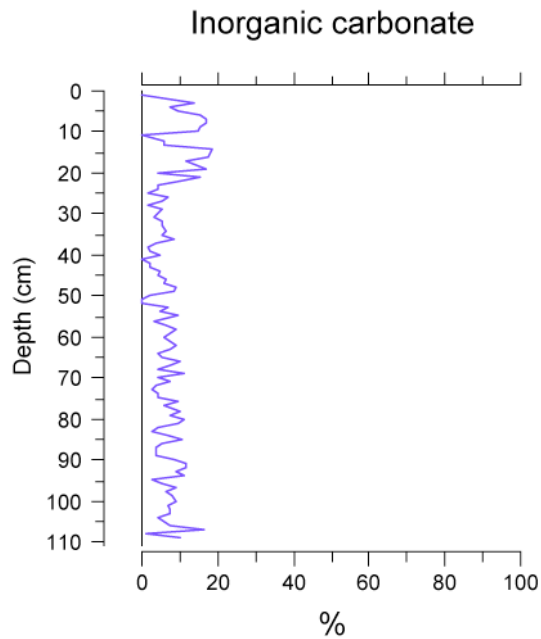


Figure 5.11: Sediment inorganic carbonate content for core LSFS

#### 5.2.1.2 Chronology

This core is the focus of a joint collaborative study with researchers from Monash University in Victoria who dated the upper sediments using  $^{210}\text{Pb}$  radionuclide data. Two samples from lower in the core were dated using AMS of the pollen fraction of the sediment.

Results of  $^{210}\text{Pb}$  analysis are given in Table 5.7 which shows the  $^{10}\text{Po}$ ,  $^{226}\text{Ra}$  and excess  $^{210}\text{Pb}$  activities of the samples and the graphical representation of these are given in Figures 5.12 to 5.15. Interpretation of these results was undertaken in consultation with analysts from the Australian Nuclear Science and Technology Organisation (ANSTO). They considered that the top 4 cm of the core indicated a mixed surface layer and, having excluded these from the interpretation, extrapolated the results from the zone beneath (11 – 31 cm) to 0 cm depth. In total, three zones of differing sedimentation rate were identified, as illustrated in Table 5.8. Using the modified Constant Initial Concentration (CIC) method described by Brugam (1978), an age of approximately 135 years was calculated for a depth of 60 – 62 cm. The three samples below this depth had essentially constant levels of  $^{210}\text{Pb}$  activity and were interpreted as being background levels beyond the scope of this dating method.

Table 5.7:  $^{210}\text{Po}$ ,  $^{226}\text{Ra}$  and excess  $^{210}\text{Pb}$  activities from core LSFS. Calculated using the modified CIC method of Brugam (1978)

Depth (cm)		ANSTO ID	Activity of Po-210 (Bq/kg)		Activity of Ra-226 (Bq/kg)		Activity of Excess Pb-210 (Bq/kg)	
from	to							
0	-	3	H663	514.85 +/- 13.49	2.09 +/- 1.72	512.76 +/- 13.60		
3	-	5	H664	601.12 +/- 17.59	2.30 +/- 1.23	598.82 +/- 17.64		
10	-	12	H665	623.86 +/- 17.40	2.96 +/- 0.75	620.91 +/- 17.42		
20	-	22	H666	207.23 +/- 8.86	4.65 +/- 0.60	202.58 +/- 8.88		
30	-	32	H667	93.00 +/- 2.91	3.17 +/- 0.51	89.83 +/- 2.95		
40	-	42	H668	96.05 +/- 3.09	4.21 +/- 0.63	91.84 +/- 3.15		
50	-	52	H669	87.28 +/- 2.83	2.80 +/- 0.48	84.47 +/- 2.87		
60	-	62	H670	28.41 +/- 1.36	2.23 +/- 0.35	26.17 +/- 1.40		
70	-	72	H671	18.59 +/- 0.95	2.72 +/- 0.41	15.87 +/- 1.03		
80	-	82	H672	16.61 +/- 0.88	2.27 +/- 0.37	14.35 +/- 0.96		
90	-	93	H673	19.30 +/- 0.90	3.27 +/- 0.52	16.03 +/- 1.04		

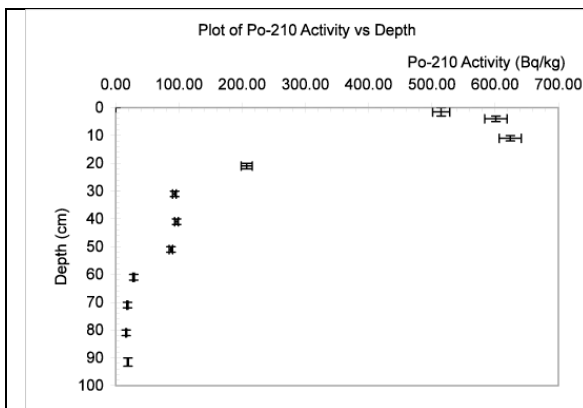


Figure 5.12: CIC-derived  $^{210}\text{Po}$  activity versus depth Lake Surprise core LSFS

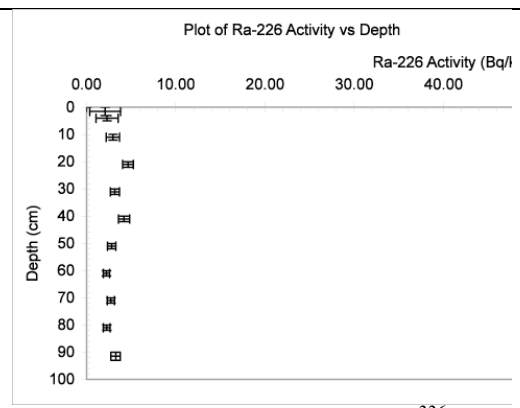


Figure 5.13: CIC-derived  $^{226}\text{Ra}$  activity versus depth Lake Surprise core LSFS

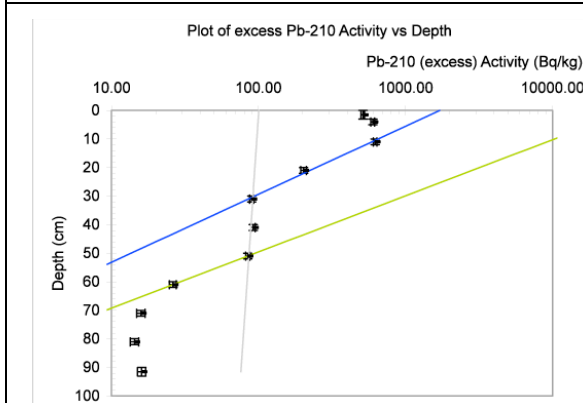


Figure 5.14: CIC-derived excess  $^{210}\text{Pb}$  activity (log-linear) versus depth Lake Surprise core LSFS

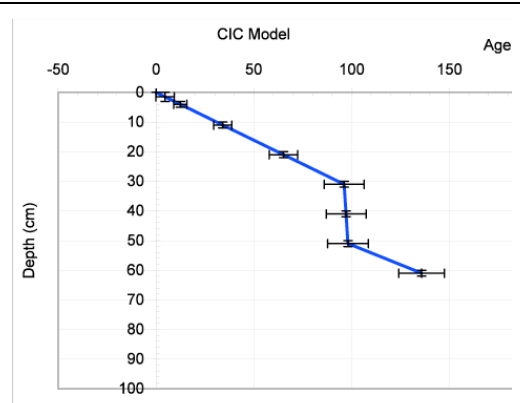


Figure 5.15: CIC-derived age versus depth Lake Surprise core LSFS

Table 5.8: Calculated sedimentation rate interpolated from  $^{210}\text{Pb}$  analysis, core LSFS, using the modified CIC method of Brugam (1978).

Depth (cm)	Sedimentation rate (cm/yr)	Correlation coefficient ( $r^2$ )
0 - 31	$0.32 \pm 0.03$	0.9917
31 - 51	$10 \pm 10$	0.5042
51 - 61	0.27	1.0000

Sediment ages were also calculated using the constant rate of supply (CRS) method of Appleby and Oldfield (1978). The resulting age determination and mass accumulation rates are presented in Table 5.9.

Table 5.9: Sample ages and mass accumulation rates determined using the CRS method of Appleby and Oldfield (1978).

ANSTO ID	Sample Depth (cm)	Dry bulk density ( $\text{g}/\text{cm}^3$ )	Calculated CRS ages (years)	CRS model Mass accumulation rates ( $\text{g}/\text{cm}^2/\text{yr}$ )
H663	0-3	0.94	$1.95 \pm 0.5$	$0.723 \pm 0.171$
H664	3-5	0.95	$5.23 \pm 0.7$	$0.713 \pm 0.089$
H665	10-12	0.95	$18.65 \pm 0.9$	$0.560 \pm 0.027$
H666	20-22	0.98	$38.28 \pm 0.9$	$0.525 \pm 0.013$
H667	30-32	1.00	$50.71 \pm 0.9$	$0.592 \pm 0.011$
H668	40-42	0.97	$62.52 \pm 0.9$	$0.638 \pm 0.010$
H669	50-52	0.98	$80.31 \pm 1.0$	$0.617 \pm 0.007$
H670	60-62	0.93	$97.28 \pm 1.0$	$0.608 \pm 0.006$
H671	70-72	0.92	$107.82 \pm 1.1$	$0.634 \pm 0.007$
H672	80-82	0.98	$118.93 \pm 1.3$	$0.655 \pm 0.007$
H673	90-93	1.02		

Two  $^{14}\text{C}$  AMS dates were undertaken on pollen extracted from the sediment at depths of 80-81 cm and 104-105 cm. The results were returned as conventional radiocarbon ages (Table 5.10), which were then calibrated (Table 5.11) using the program Calib, version 5.0.1. Calibration plots are presented in Appendix 6.



Table 5.10: Results of  $^{14}\text{C}$  AMS analysis of samples from core LSFS. Note: \* - denotes assumed  $\delta^{13}\text{C}$  as measured data is unavailable from laboratory.

LAB. CODE	Sample Depth	$\delta(^{13}\text{C})$ per mil	% Modern Carbon		Conventional $^{14}\text{C}$ Age	
			pMC	1 $\sigma$ error	Yrs BP	1 $\sigma$ error
OZI015	80-81cm	-25.0*	93.11	0.67	570	60
OZI016	104-105cm	-25.0*	91.70	0.82	700	80

Table 5.11: Results of radiocarbon age calibration (undertaken using the SHCal04 data set of McCormac et al. (2004) in Calib 5.0.1)

Sample	$^{14}\text{C}$ Age	One sigma age ranges	Relative Area under probability distribution	Two Sigma age ranges	Relative Area under probability distribution	AGE cal yr B.P	AGE AD/BC
OZI015	570 $\pm$ 60	516-553 BP	1	501-563 BP 602-628 BP	0.888 0.111	532	1418
OZI016	700 $\pm$ 80	560-608 BP 625-664 BP	0.55 0.44	538-688 BP 702-719 BP	0.97 0.02	613	1337

A graphical illustration of average sedimentation rates, as calculated from the CIC-derived and CRS-derived age/depth profiles, are presented in Figures 5.16 and 5.17 respectively. Both figures also include the calibrated  $^{14}\text{C}$ -AMS dates and demonstrate a notable age difference. This is discussed further in Section 6.1.

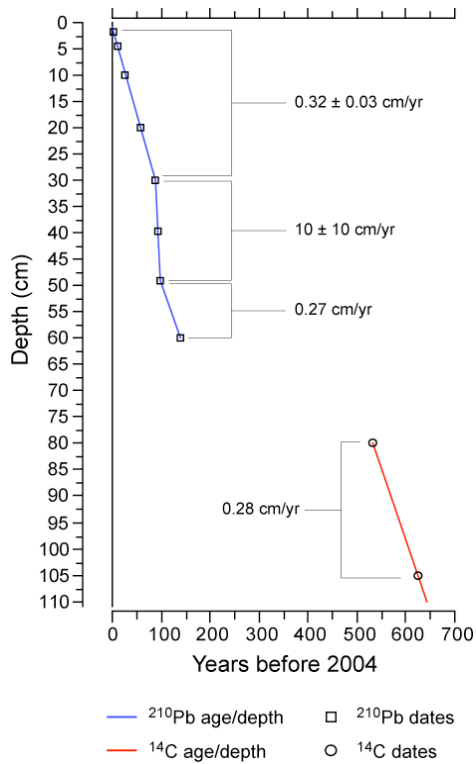


Figure 5.16: Core LSFs calculated average sedimentation rates from  $^{210}\text{Pb}$  age/depth profile, as derived from the modified constant initial concentration method (Brugam, 1978), and calibrated  $^{14}\text{C}$ -AMS dates.

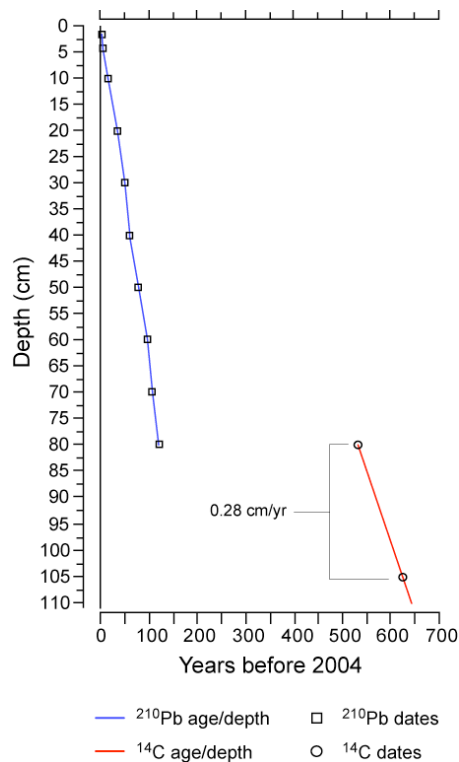


Figure 5.17: Core LSFs  $^{210}\text{Pb}$  age/depth profile, as derived from the constant rate of supply method (Appleby and Oldfield, 1978), and calibrated  $^{14}\text{C}$ -AMS dates. For mass accumulation rates, see Table 5.9.

### 5.2.1.3 Fossil diatom analysis

The fossil diatom record from Core LSFS is presented in Figure 5.18. Species counts are presented as percentage of total number of valves counted and all species which constituted a minimum of 5% of the count in at least one sample are included.

#### *Zone LSFS-4 (110 cm – 79 cm)*

This zone encompasses samples from the base of the core (110 cm) to a depth of 77 cm. The percentage of planktonic species drops to a low of 24% at 96 cm before gradually increasing to a high as 56.5% at 80 cm. Of the planktonic species, there is a gradual decline in the abundance of *Cyclotella meneghiniana* moving upwards through the zone while *Cyclostephanos dubius* maintains relatively constant levels. *Discostella pseudostelligera*, which is almost absent throughout the remainder of the record, attains its maximum abundance in this zone, consisting of 33.2% of the total valve count at its peak at 82 cm. Similarly, *Discostella stelligera*, which attains a maximum abundance of 6.9% at the zone boundary of 78 cm, is present in significant numbers in this zone only.

The majority of the benthic taxa exhibit relatively constant levels of abundance with minor variations throughout the zone. The members of the ‘small *Fragilaria*’ group, namely *Staurosira elliptica*, *Staurosirella pinnata* and *Staurosirella leptostauron*, all peak around 94-96 cm. *Synedra acus*, which is relatively constant throughout the majority of the zone, rapidly diminishes towards the top of the zone and only accounts for 0.4% of the species assemblage at 80 cm. Valve density within this zone is relatively constant, though does demonstrate some minor fluctuations across the middle of the zone.

#### *Zone LSFS-3 (78 cm – 55 cm)*

The levels of planktonic taxa in this zone are reasonably constant around 50% of the total count. By far, the majority of this contribution is provided by *C. dubius* which dominates the planktonic count. *Cyclotella meneghiniana* has a constantly low representation, while *D. pseudostelligera* and *D. stelligera* rapidly diminish at the beginning of the zone. *Stephanodiscus hantzschii* has low, but consistent, levels before disappearing from the record at a depth of 56 cm. Again, most of the benthic taxa show relatively constant levels throughout the zone. *Staurosira elliptica* and *Staurosirella pinnata* gradually increase towards the top of the zone while *Synedra acus* shows

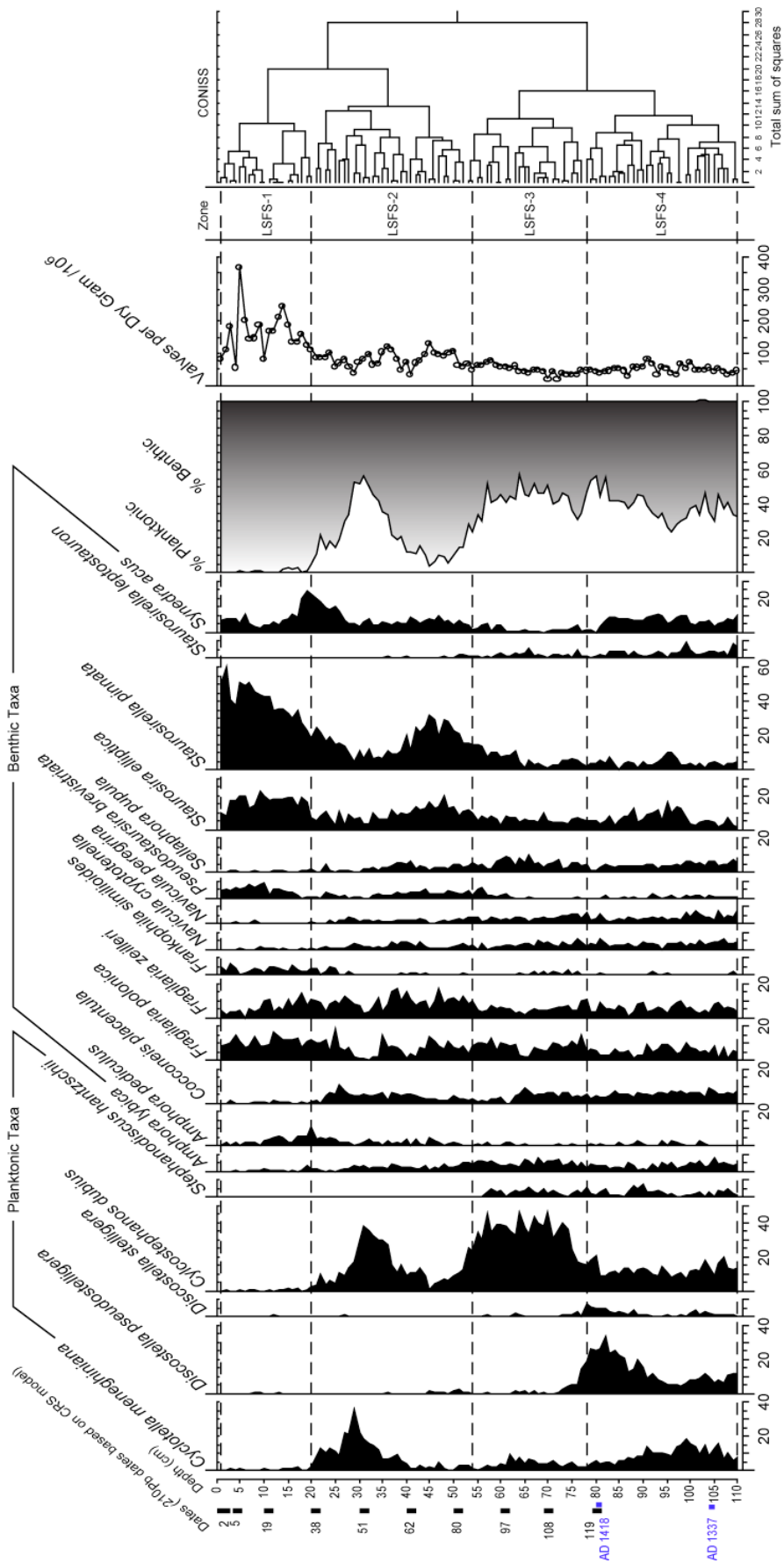
consistently low levels. Valve density was relatively constant across the zone with only minor fluctuations.

*Zone LSFS-2 (54 cm – 20 cm)*

The key feature within this zone is the rapid alternation from a community dominated by benthic taxa to one dominated by planktonic taxa. At 45 cm, planktonic taxa only account for 4.1% of the total count before rising to a peak of 56.4% of the count at 31 cm, before again falling back to 4.7% at the upper zone boundary at 19 cm. Only two planktonic species account for this increase, *Cyclotella meneghiniana* and *Cyclostephanos dubius*, with the peak in *C. dubius* preceding that of *Cyclotella meneghiniana* by 2 cm.

The benthic taxa are dominated by the small *Fragilaria* group of *Staurosira elliptica* and *Staurosirella pinnata* and, while *Staurosira elliptica* is the slightly less dominant of the two, their pattern of abundance mirror each other with a peak at the base of the zone followed by a decrease in abundance coinciding with the dominance of the planktonic taxa. This is followed by a steady increase in *Staurosirella pinnata* toward the top of the zone, which is also evident in *Synedra acus* which has re-emerged in prominence and constitutes 19.7% at 21 cm. The abundance of *Amphora libyca*, *Cocconeis placentula*, *Navicula cryptotenella* and *Sellaphora pupula* all decrease toward the top of the zone and remain infrequent for the remainder of the core. *Fragilaria zeilleri* and *Fragilaria polonica* maintain a relatively constant presence with the latter showing a marked decrease coinciding with the planktonic dominance, a trait not evident in previous zones. Counts of both *Amphora pediculus* and *Pseudostaurosira brevistriata* become more prominent within this zone having only shown low values previously. Having been relatively constant in the preceding zones, the density of diatom valves is slightly greater in this zone and displays an increased level of variability.

Figure 5.18: Core LSFS diatom stratigraphy showing all taxa with a relative abundance of 5% or more



### *Zone LSFS-1 (19 cm – 1 cm)*

This zone represents the uppermost 20 cm of the core, including the sediment/water interface. There is a marked dominance of benthic taxa which represent more than 99% of the count in the majority of samples. The greatest contributors to this are, once again, the small *Fragilaria* group, yet, while *Staurosira elliptica* remains relatively constant at around 17% of the count, *Staurosirella pinnata* maintains a steady increase across the zone to a maximum value of 59.9% at a depth of 2 cm. The zone boundary at 20 cm coincides with a peak in proportions of *A. pediculus* (10% of the count) while the maximum levels of *Synedra acus* (24%) occur at 19 cm. The larger *Fragilaria*-type species of *Frankophila similioides* and *Pseudostaurosira brevistriata* also attain their maximum levels of representation across this zone. The number of valves per dry gram of sediment is markedly elevated in this zone, reaching a maximum of  $369 \times 10^6$  at a depth of 5 cm, and showing notable fluctuations.

## 5.2.2 Core LST1

### *5.2.2.2 Sediment lithology*

The lithology of Core LST1 is illustrated in Figure 5.19. The most notable features of the core are the numerous bands of lighter sediment which often have a sharply delineated boundary with adjoining sedimentary units. These formed the basis of the location of radiocarbon dates and are in contrast to the majority of the upper section of the core which is a homogenous dark brown lake mud.

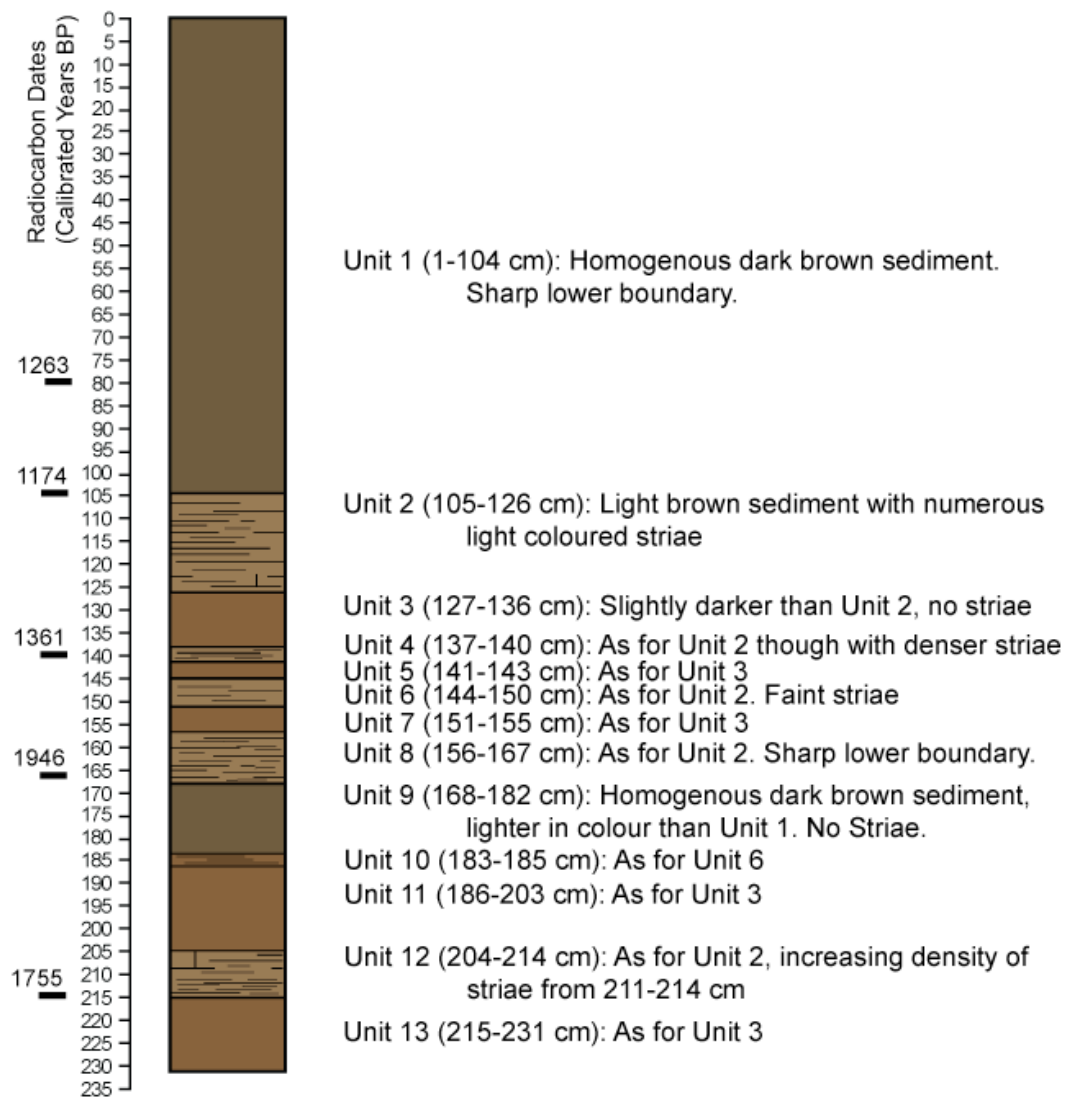


Figure 5.19: Sediment lithology for core LST1

#### 5.2.2.1 Sediment analysis

As was the case for Core LSFS, contiguous samples were taken for sedimentary analysis of bulk density, moisture content, organic content and inorganic carbonate content, at 1 cm intervals. This core measures 231 cm in length and demonstrates marked fluctuations in some of the variables measured. Results are presented in Figures 5.20 to 5.23.

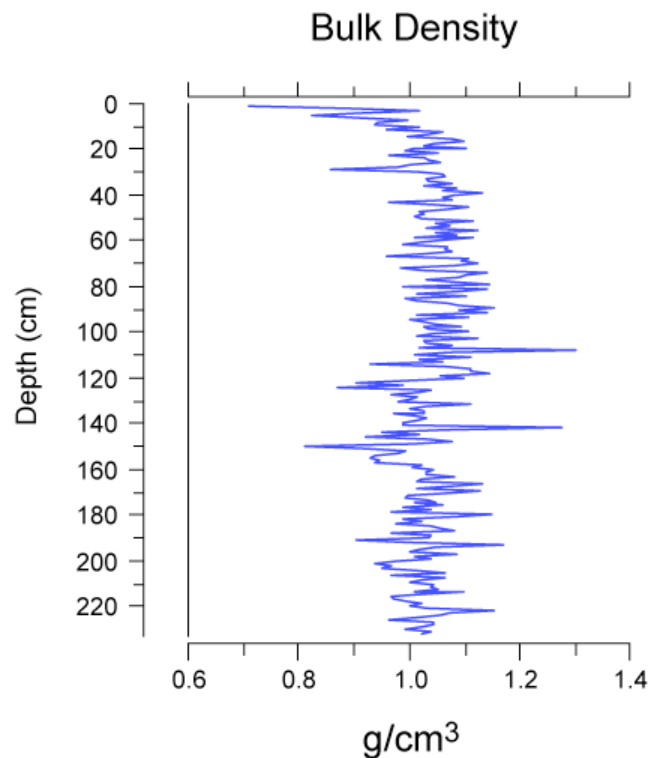


Figure 5.20: Bulk density calculations for core LST1

From the base of the core at 231 cm to a depth of 161 cm, there is some variability within bulk density measurements, though the overall trend across this portion of the core is relatively stable and averages  $1.03 \text{ g/cm}^3$ . Over the next 55 cm of core, from 160 cm to 105 cm, there are large fluctuations, often from one centimetre to the next. Within this portion of the record, from 160 cm to 120 cm, there is a pattern of generally lower bulk density than the remainder of the record, though obvious and occasionally large fluctuations still exist. Apart from a gradual decrease over the top 20 cm of the core, the bulk density of the upper metre of sediment remains reasonably stable with an average of  $1.04 \text{ g/cm}^3$ .

The sedimentary moisture content of core LST1 shows clear fluctuations (Figure 5.21). From the base of the core to a depth of approximately 214 cm, moisture content is generally elevated and averages around 89%. Thereafter, there is a sharp decline in moisture content to 86.5% at 112 cm, before a more gradual a more gradual trend of decreasing sedimentary moisture until 160 cm. The 56 cm of core from 160 cm to 104 cm is the most variable period of the record. There is a sudden and dramatic increase in both the frequency and amplitude of variability, with the largest change



occurring over a 4 cm interval between 153 cm (89.7%) and 149 cm (83.1%). From 103 cm to 27 cm, variability is markedly decreased with moisture content remaining relatively stable, with small fluctuations between 87% and 90%. From a high of 87.8% at 27 cm, there is a rapid decline in sediment moisture content to 85.7% at 22 cm, before a sudden increase to 89.5% at 16 cm. Moisture content then falls away again, to a low of 82% at a depth of 5 cm.

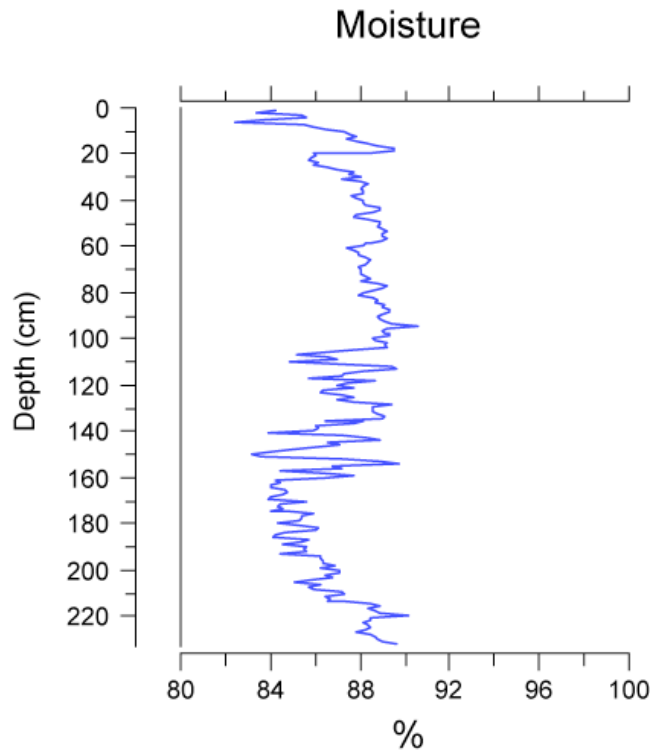


Figure 5.21: Moisture content calculations for core LST1

The organic content of the sediment shows a similar pattern to that of the moisture content (Figure 5.22). From the base of the core to 160 cm, there general trend indicates low organic content, though notable variability is evident. From 160 cm to 103 cm there is, again, a sudden shift to a sustained period of increased variability with rapid fluctuations between samples of high and low organic content. The top metre is relatively stable showing slight fluctuations over a background of approximately 60% organic content.

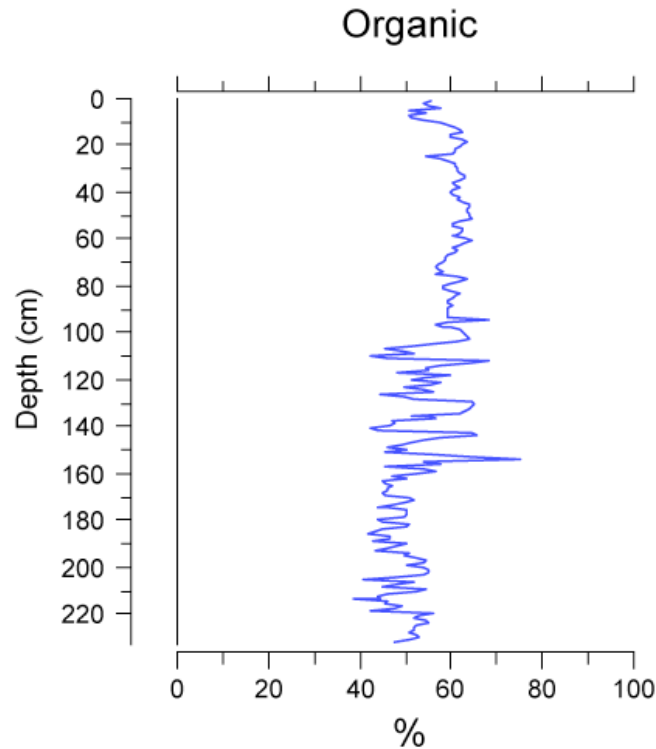


Figure 5.22: Sediment organic content for core LST1

The inorganic carbonate content of the sediment similarly reflects the results of both the moisture and organic content (Figure 5.23). From the base of the core to 217 cm, carbonate content remains relatively stable, averaging 8.1%. From this point, until 168 cm, variability increases dramatically, though the background trend indicates gradually increasing carbonate content. From 167 cm until 105 cm, the amplitude of variability increases further and though the frequency of variation is slightly less than the previous 49 cm, there are sudden shifts evident from samples with high carbonate content to those with low carbonate content and vice versa. The uppermost metre indicates a period of relative stability which is most apparent from 104 cm to 70 cm.

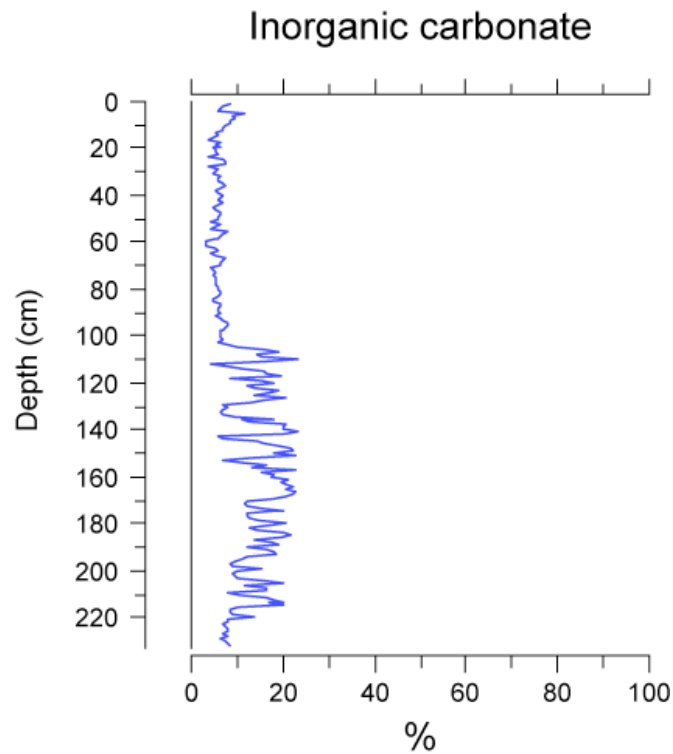


Figure 5.23: Sediment inorganic carbonate content for core LSFS

### 5.2.2.3 Chronology

Dating of core LST1 comprised of five  $^{14}\text{C}$  AMS dates on the pollen fraction of the sediment. The location of dates was guided, in part, by noted changes in the sediment lithology (see above). No  $^{210}\text{Pb}$  dates were undertaken as it was assumed that the uppermost sediments would be captured by the frozen spade core LSFS. Results of  $^{14}\text{C}$  AMS analysis are presented in Table 5.12, while the results of age calibration are presented in Table 5.12. Calibration plots are presented in Appendix 7.

Table 5.12: Results of  $^{14}\text{C}$  AMS analysis of samples from core LST1. Note: \* - denotes assumed  $\delta^{13}\text{C}$  as measured data is unavailable from laboratory.

LAB. CODE	Sample Depth	$\delta^{13}\text{C}$ per mil	Percent Modern Carbon		Conventional $^{14}\text{C}$ Age	
			pMC	1 $\sigma$ error	Yrs BP	1 $\sigma$ error
OZI591	80-81cm	-27.7	84.09	0.56	1390	60
OZI017	104-105cm	-26.0	85.02	0.48	1300	50
OZI019	141-142cm	-25.0*	83.07	0.99	1490	100
OZI020	167-168cm	-25.9	77.67	0.80	2030	90
OZI021	214-215cm	-26.6	79.29	0.55	1860	60

Table 5.13: Results of radiocarbon age calibration (undertaken using the SHCal04 data set of McCormac et al. (2004) in Calib 5.0.1)

Sample	<sup>14</sup> C Age	One Sigma age ranges	Relative area under probability distribution	Two Sigma age ranges	Relative area under probability distribution	AGE cal.yr. B.P.	AGE AD/BC
<b>OZI591</b>	1390±60	1187-1201 1257-1299	0.177 0.822	1179-1214 1221-1306	0.206 0.794	1263	687
<b>OZI017</b>	1300±50	1143-1159 1169-1189 1199-1260	0.131 0.209 0.660	1082-266	1	1174	776
<b>OZI019</b>	1490±100	1265-1418 1466-1490 1498-1508	0.894 0.073 0.032	1140-1160 1168-1554	0.009 0.991	1361	589
<b>OZI020</b>	2030±90	1825-1853 1857-2000	0.116 0.884	1737-1763 1769-2117	0.025 0.975	1943	7
<b>OZI021</b>	1860±60	1701-1744 1750-1813	0.446 0.554	1615-1678 1685-1825 1853-1858	0.161 0.832 0.005	1755	195

Due to the ‘reversal’ of ages for samples at 104-105 cm and 167-168 cm, several possible scenarios exist for calculating average sedimentation rates. These are discussed further in Section 6.1, in the context of formulating a final chronology for core LST1, and are therefore not presented here.

#### 5.2.2.4 Fossil diatom analysis

In comparison to core LSFS, a larger number of species attain a relative abundance of greater than 5% in any one sample. Therefore, in order to more clearly illustrate the results, two diagrams are presented. Figure 5.24 illustrates all species which attained a relative abundance of between 5% and 15% in at least one sample, and includes sediment lithology, valve density and CONISS dendrogram. Figure 5.25 includes all species which attained a relative abundance of greater than 15% a single sample.

The main feature of the fossil diatom counts from Core LST1 are the rapid fluctuations in the abundance of several species, in particular the planktonic taxa, though also some benthic species as well. Six main zones have been identified, each will be described below in chronological order.

Figure 5.24: Stratigraphic diatom diagram from core LST1 showing all species with a relative abundance of 5-15% in one sample

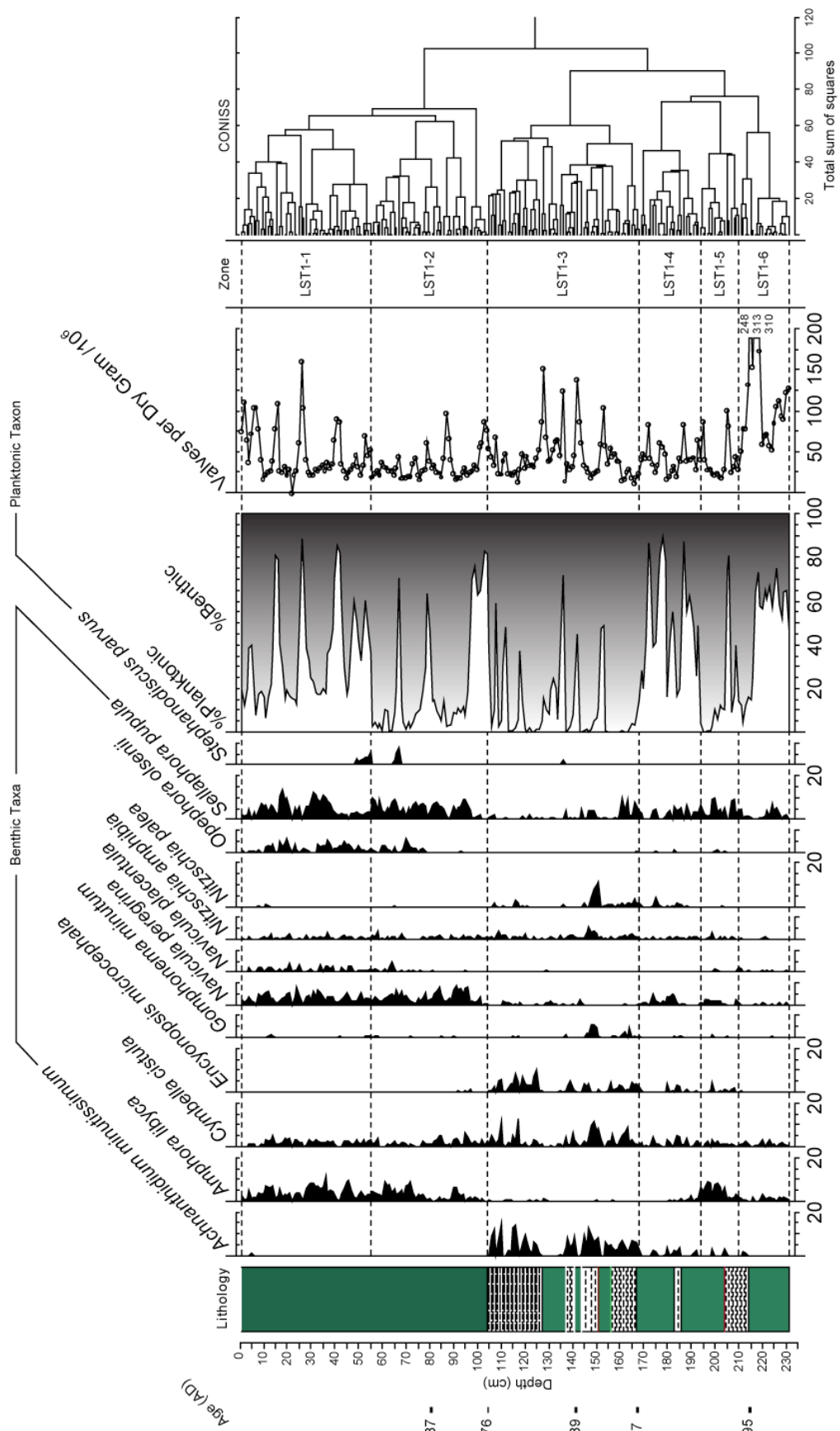
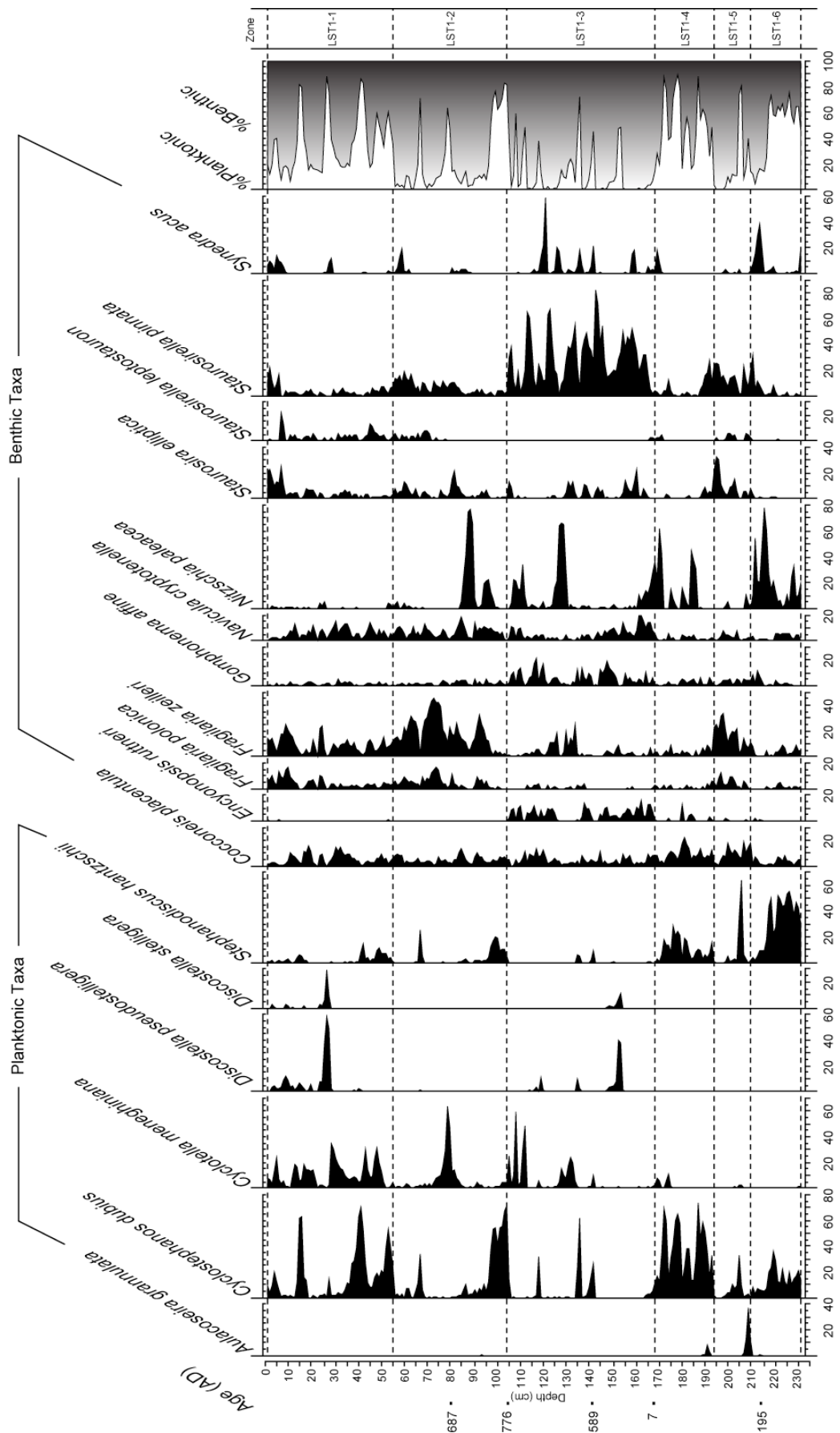


Figure 5.25: Stratigraphic diatom diagram from core LST1 showing all species with a relative abundance of 15% or more in one sample.



#### *Zone LST1-6 (231-210 cm)*

This zone is dominated, in the first instance, by the planktonic taxa *Cyclostephanos dubius* and *Stephanodiscus hantzschii*, which account for roughly 60% of the total count. Benthic taxa then dominate beginning with *Nitzschia paleacea*, which attains 60.7% and 78.3% of the count at 216 cm and 215 cm respectively, with a concurrent rise in valve density. The boundary between Unit 13 and Unit 12 in the sediment lithology occurs at 214 cm and the period between 214 cm and 211 cm shows a greater concentration of bands in the sediment. This period corresponds to a sudden rise in *Synedra acus* (38% at 214 cm and 29% at 213 cm) which is followed by another sharp peak in *N. paleacea* at 212 cm (54.7%) and an increase in *Staurosirella pinnata* at 211 cm (32%).

#### *Zone LST1-5 (210-194 cm)*

The zone boundary is marked by a sudden peak in *Aulacoseira granulata* which, having been absent at 211 cm, dominates the count at 209 cm with 37.25%. The peak is short-lived (210 - 207 cm) and, by 207 cm, *Cocconeis placentula* (18.8%) and *S. pinnata* (23.2%) have become dominant. There follows a brief return to domination of planktonic forms at 206 cm and 205 cm as *Stephanodiscus hantzschii* (64% and 36.6%) and *Cyclostephanos dubius* (8.9% and 32.4%) dominate. At 204 cm there is a sudden return to dominance of the benthic species which total 89% of the species counted. This corresponds to the lithological boundary between Unit 12 and Unit 11. The remainder of the zone is dominated by non-planktonic taxa with *Cocconeis placentula*, *Fragilaria polonica*, *F. zeilleri*, *Staurosirella pinnata*, *Amphora libyca* and *Sellaphora pupula* all maintaining relatively constant degrees of representation, while *Staurosira elliptica* gradually increases in abundance across the zone.

#### *Zone LST1-4 (194-168 cm)*

The base of the zone is marked by the rapid decrease in the proportions of *Staurosirella pinnata* and *Staurosira elliptica* followed by a small peak in *Aulacoseira granulata*. *Cocconeis placentula* maintains its presence, though the zone is dominated by *Stephanodiscus hantzschii* and *Cyclostephanos dubius*, the latter of which exhibits a number of rapid fluctuations. The stratigraphic unit 10 (183 – 185 cm) corresponds with a sudden peak in the dominance of benthic taxa, primarily driven by *Nitzschia paleacea*. After this, peaks in *N. paleacea* alternate with those of *C. dubius*. This is most apparent

toward the top of the zone where the representation of *C. dubius* falls from 68.8% to 13% between 171 cm and 169 cm, and is replaced by *N. paleacea*, which rises from 0% to 62.3% over the same period.

#### *Zone LST1-3 (168-104 cm)*

As illustrated by the CONISS dendrogram, this zone is the most dissimilar, in terms of species assemblages, of all zones identified. Though the zone boundaries are derived from the diatom data, they correspond with the boundaries of sedimentary units, an attribute not apparent in the zones described above. The zone also contains seven of the 13 lithological units outlined in Figure 5.19 above.

In terms of diatoms, the zone is clearly dominated by benthic taxa, particularly *S. pinnata* which attains a maximum representation of 82.5% of the total count at a depth of 142 cm. Several other benthic taxa also attain their highest representation in the record, including *Achnanthydium minutissimum*, *Encyonopsis ruttneri*, *Encyonopsis microcephala*, *Gomphonema affine* and *Cymbella cistula*. Small peaks of *Gomphonema minutum*, *Nitzschia amphibia* and *Nitzschia palea* occur simultaneously, suggesting some level of ecological affinity.

The first appearance of *Discostella pseudostelligera* and *Discostella stelligera* in the record occurs at 153 cm and corresponds with Unit 7 in the sediment lithology, which consisted of homogenous lake mud with no banding present. Their representation rapidly diminishes in sedimentary Unit 6 (144 – 150 cm), in which faint banding is present. Instead, this portion of the record is associated with (relatively) large peaks in *G. affine*, *G. minutum*, *A. minutissimum*, *C. cistula* and *N. palea*. An increase in the representation of planktonic taxa, from 0% to 45% of the count, correlates with sedimentary Unit 5 (141 – 143 cm), which is, again, homogenous lake mud with no banding present. Sedimentary Unit 4 (137 – 140 cm) consists of light brown sediment with dense striae and corresponds with a return of non-planktonic taxa, primarily *Staurosirella pinnata*, *Staurosira elliptica*, *Gomphonema affine* and *Encyonopsis ruttneri*, with planktonic taxa accounting for only 0.5% of the count through this portion of the record. A sudden increase in the proportions of *Cyclotella dubius*, from 0.5% at 137 cm to 62% at 136 cm, aligns with the lower boundary of Unit 3, though the peak is short-lived and is rapidly replaced by *Cyclotella meneghiniana*, *Fragilaria zeilleri* and *N. paleacea*. The next 21 cm of the zone corresponds to Unit 2 and is



characterised by large and rapid fluctuations in proportions of *Staurosirella pinnata*, *Synedra acus*, *C. meneghiniana* and, to a lesser extent, *Cyclostephanos dubius* and *N. paleacea*.

#### *Zone LST1-2 (104-55 cm)*

Primarily driven by a very sharp increase in the representation of *C. dubius* and *Stephanodiscus hantzschii*, planktonic taxa dominate the first 7 cm of this zone. Their values rapidly diminish and are replaced by peaks of *N. paleacea* (76.7% at 88 cm) and greater proportions of *Fragilaria zeilleri* and *F. polonica*. Apart from a large spike in *Cyclotella meneghiniana* (63.5% at 79 cm), and peaks in *Cyclostephanos dubius* and *S. hantzschii* at 67 cm, the zone is dominated by benthic taxa. Of these, *Amphora libyca*, *Navicula peregrina* and *Sellaphora pupula* all become more prevalent and maintain a presence to the top of the core. Proportions of the dominant taxon in the previous zone, *Staurosirella pinnata*, fall rapidly at the base of the zone before gradually increasing toward the top of the zone.

#### *Zone LST1-1 (55-1 cm)*

The beginning of the zone is marked by a sudden increase in proportions of planktonic taxa in the record. This is primarily driven by an increase in numbers of *C. dubius* and *Stephanodiscus hantzschii*, though these are soon replaced by *Cyclotella meneghiniana*. The rapid increases and decreases in planktonic taxa, that was evident in previous zones, is continued here with sharp peaks in *Cyclostephanos dubius*, *Discostella pseudostelligera* and *D. stelligera*. The latter two species attain their highest representation in the record at 26 cm, with 57.8% and 39.6% of the count respectively. In general though, proportions of planktonic taxa gradually decrease toward the top of the zone and benthic taxa become more prominent.

Despite the increasing dominance of benthic taxa, no individual species dominates the count with numerous species contributing to the overall benthic proportions. *Cocconeis placentula*, *Amphora libyca*, *Cymbella cistula*, *Navicula peregrina*, *Opephora olsenii* and *Sellaphora pupula* all maintain fluctuating, but low, numbers. The representation of *Fragilaria polonica* and *F. zeilleri*, are also relatively constant, though they increase in the upper 13 cm of the zone, as does *Staurosira elliptica*, *Staurosirella pinnata* and *Synedra acus*.

### 5.3 Correlation of Lake Surprise cores

Where multiple sediment cores are taken from the same site, some form of correlation is required. Multiple cores may be taken to gain a greater understanding of the limnology of the water body (e.g., Anderson, 1990; Leahy *et al.*, 2005) or, as was the case in Lake Surprise, to extend the chronological extent of the sedimentary record. Different circumstances require different coring methods and this was the case with the Lake Surprise record where the sediment/water interface was large (approximately 50 cm) and the moisture content of that portion of the sediment record was very high (max: 98.9%; average: 95.5% in core LSFS). In a situation such as this, most coring methods would result in mixing of the upper sediments and loss of stratigraphic integrity. Therefore, to capture this component of the Lake Surprise record, a frozen spade corer was employed (Neale and Walker, 1996). Due to the design of the corer though, penetration into the sediments is limited and a different coring method was required to capture the longer record, commensurate with the aims of the project. In order to formulate a continuous record, these two cores need to be correlated by determining the degree to which they overlap, thus enabling the two records to be merged.

Correlation of sedimentary archives can be undertaken in several ways. Commonly, this involves comparisons between magnetic susceptibility (Mooney, 1997; Bertrand *et al.*, 2005), lithostratigraphy (Stanley and De Deckker, 2002; Bertrand *et al.*, 2008), biostratigraphy (Anderson, 1990), sediment analysis (Halfman *et al.*, 1994) or a combination of these methods (Leahy *et al.*, 2005). Correlating the data can also be undertaken statistically using methods such as sequence slotting (Thompson and Clark, 1989). Often however, it is simply undertaken by comparing the relevant graphs and aligning them visually, a task that is simplified where significant changes or obvious markers are evident.

At Lake Surprise, the initial aim was to correlate the cores using a combination of sediment stratigraphy, sediment analysis, the biostratigraphic changes evident in the fossil diatom record and radiocarbon dating. However, the captured sediment in the top metre of core LST1, and the entire length of core LSFS, proved to be homogenous with no obvious stratigraphic markers evident, thereby precluding it as a useful means of correlation. Despite this, the homogenous nature of the sediment in both cores gave

weight to the assumption that the upper 110 cm of core LSFS would comfortably overlap with the top 100 cm of core LST1.

Comparisons between the results of the sedimentary analysis though, also proved inconclusive, with no matching point of correlation apparent in the sediment analyses. Bulk density analysis (Figure 5.26) indicated a broad agreement between the downward trend in the bottom 10 cm of LSFS and the top of LST1, though the degree of variability between samples differed. Similarly, comparisons of sedimentary organic content (Figure 5.27) also suggested a general agreement, though, in this instance, an approximate 20 cm overlap seemed most warranted. Analysis of the inorganic carbonate content of the cores (Figure 5.28) suggests that an approximate 15 cm overlap, while results of sedimentary moisture content analysis (Figure 5.29) showed no obvious point of correlation at all.

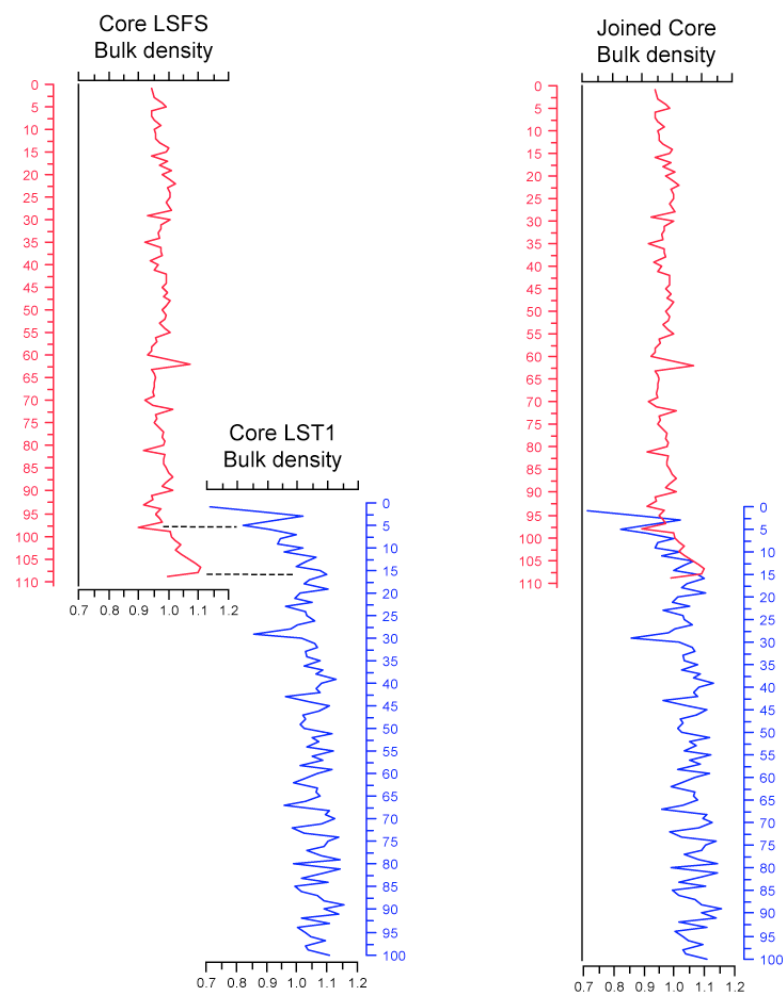


Figure 5.26: A comparison of sedimentary bulk density results from core LSFS (in blue) and the top 100 cm of core LST1 (red), indicating possible overlap.

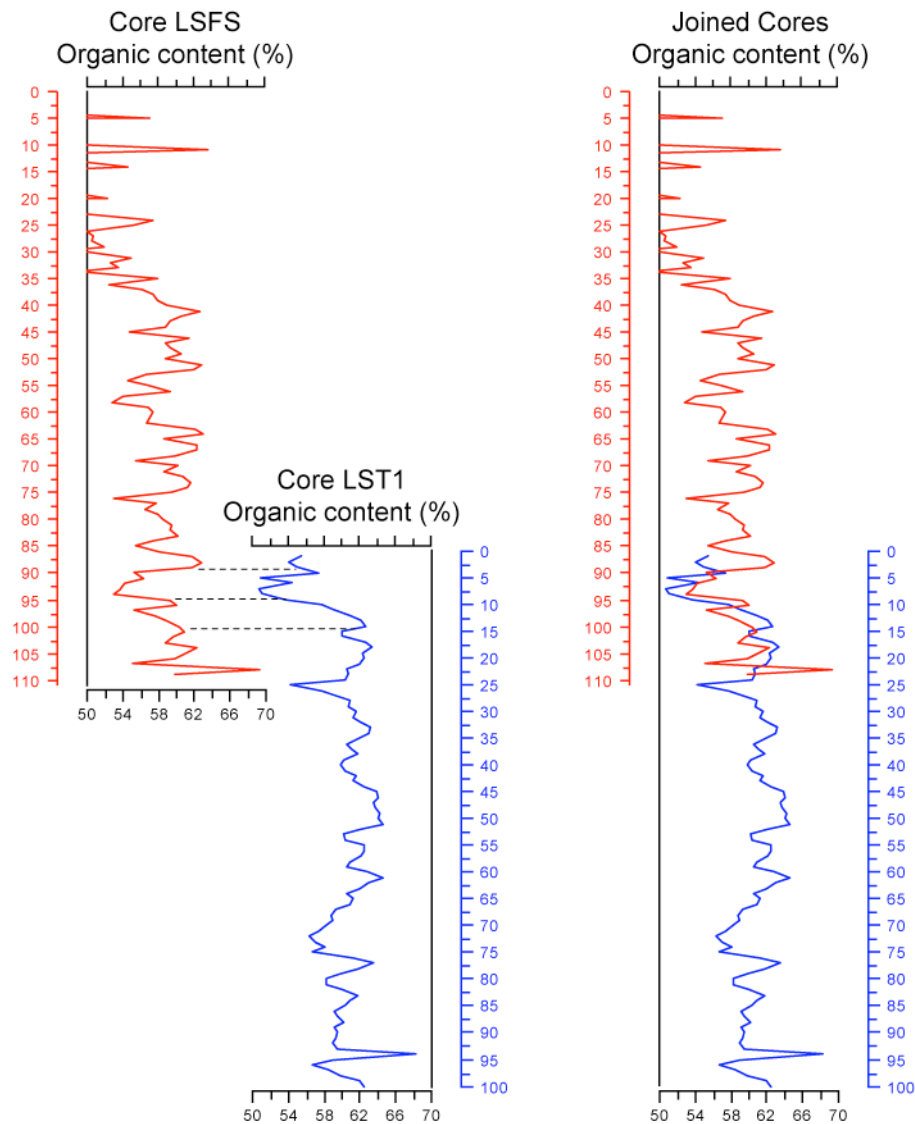


Figure 5.27: A comparison of sedimentary organic content results from core LSFS (in blue) and the top 100 cm of core LST1 (red), indicating possible overlap.

Comparing sedimentary analysis results therefore proved inconclusive, though it could be argued that correlating cores on the basis of loss on ignition results alone should be undertaken with caution in the first place. Heiri *et al.* (2001) demonstrated that the potential for errors in weight loss calculations is significant and can be dependent upon factors such as sample weight, exposure time and even factors as simple as the location of samples in the furnace. These errors were generally in the range of a few percent of dry weight lost, suggesting that their utility as a tool for core correlation is highly limited in situations where no major, or sustained, shift in measured variables is noted.

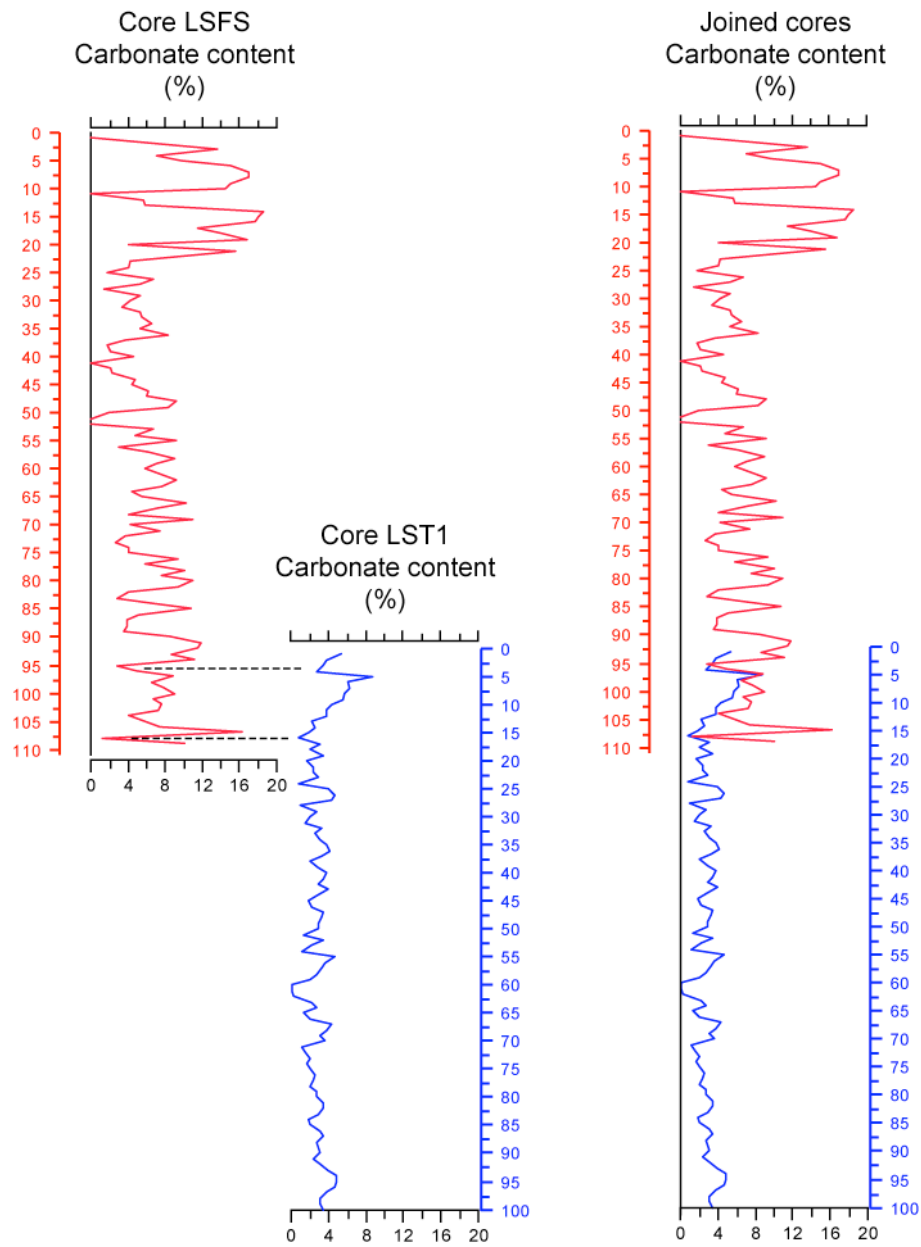


Figure 5.28: A comparison of sedimentary carbonate content results from core LSFS (in blue) and the top 100 cm of core LST1 (red), indicating possible overlap.

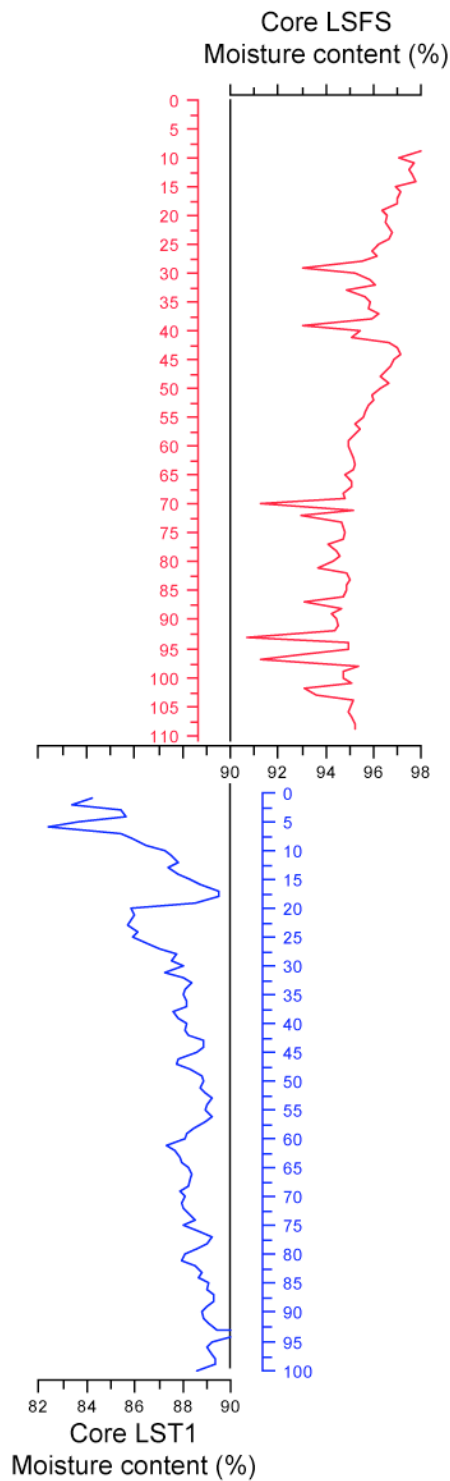


Figure 5.29: A comparison of sedimentary moisture content results from core LSFS (in blue) and the top 100 cm of core LST1 (red)

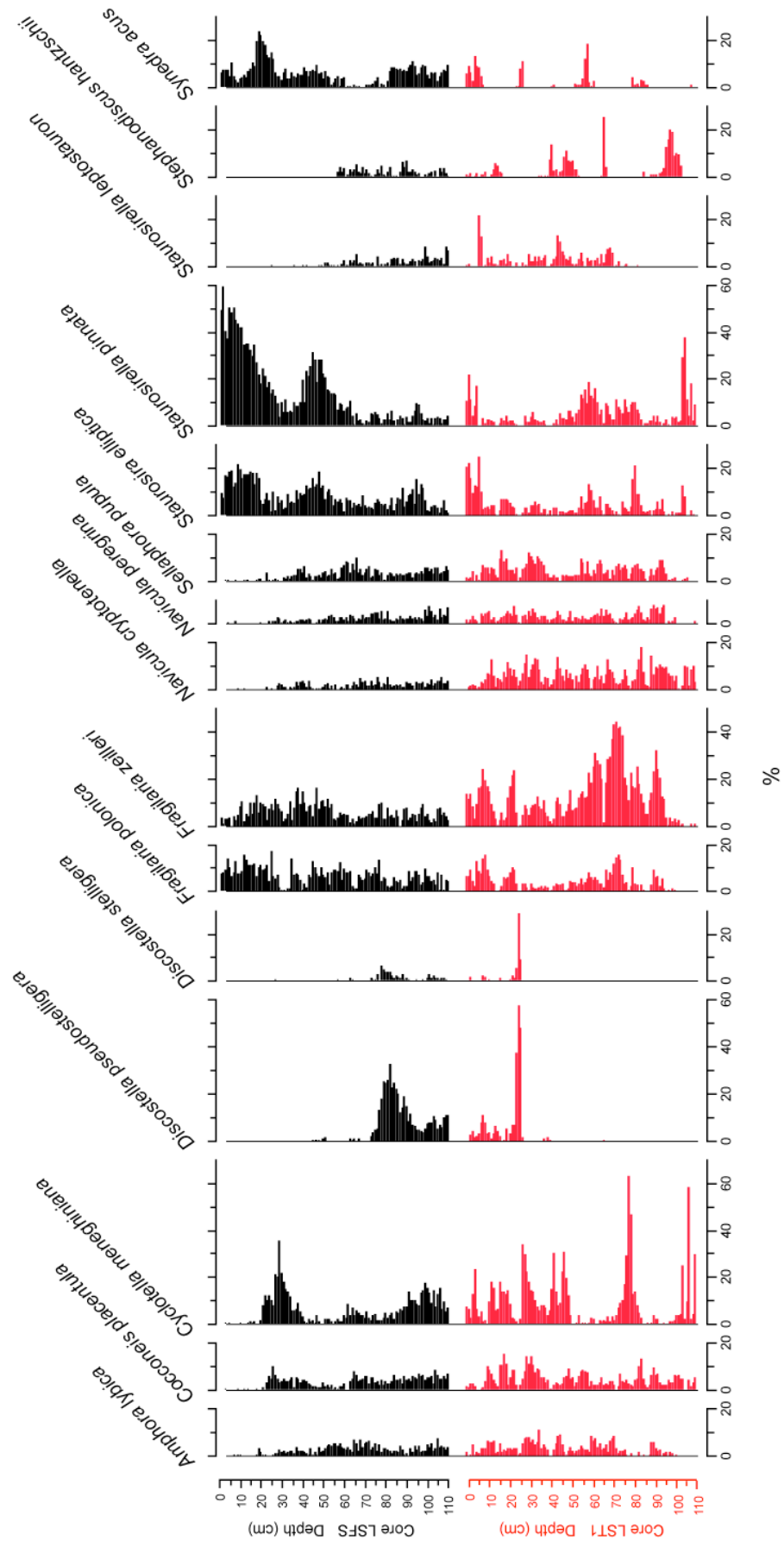
Comparison of the relative abundance of the main diatom species in both cores is similarly inconclusive indicating that there is no degree of core overlap (Figure 5.30). In general, species abundance data from LST1 shows a greater degree of centimetre to centimetre variability than what is evident in LSFS. For example, *Amphora libyca*,

*Cocconeis placentula* and *Cyclotella meneghiniana* all show constant, or gently fluctuating, levels towards the base of LSFS, but show more variable levels at the top of LST1. Similarly, peaks in *Fragilaria zeilleri*, *Staurosira elliptica* and *Staurosirella pinnata* at the top of LST1 are not evident in the LSFS record.

Even though this indicates that the cores do not overlap, it does suggest that the degree to which the cores 'miss' each other is only slight. For example, the relative abundances of *Discostella pseudostelligera*, show a gradual decrease towards the base of LSFS, a trend that is continued at the top of LST1. While over the same period, *Discostella stelligera* is present in very low numbers in both cores and *Stephanodiscus hantzschii* shows a similar pattern. The record of *Synedra acus* also supports this theory. This species only appears sporadically in the LST1 record, though maintains a constant presence over the top 10 cm of LST1, a trend which continues into core LSFS.

An examination of the chronology of the cores supports the notion that the cores do not overlap and also provides an indication of the width of the gap between them. The two calibrated, radiocarbon dates from LSFS (532 years at 80 cm and 613 years at 104 cm) indicate that each centimetre of the intervening sediment represents approximately 3.52 years. Commencing at the calibrated age of the sample at 104 cm of 613 years, and extrapolating at 3.52 years/cm to the base of the core at 110 cm, a basal age of approximately 634 years is calculated. In comparison, the calculated sedimentation rate between the upper two dates of LST1 (1174 years at 104 cm and 1361 years at 141 cm) indicates that each centimetre equates to 5.05 years which, when extrapolated to the top of the core, provides a surface age of 648 years. This suggests that the gap between the base of LSFS and the surface of LST1 is approximately 14 years. Based on the LSFS sedimentation rate of 3.52 years/cm, this equates to a 3.97 cm gap, however, based on the LST1 sedimentation rate of 5.05 years/cm, it suggests a gap of 2.77 centimetres.

Figure 5.30: Comparison of the relative abundance of the main diatom species in cores LSFS and the top 110 cm of LST1.





It is possible that the actual gap between the cores is greater than this. The calculations used here assume that a constant sedimentation rate exists between all dates and that this extends to the base (LSFS), or surface (LST1), of the core. In the case of core LSFS, this encapsulates 30 cm of core and approximately 102 years. In the case of LST1 though, this covers 141 cm of sediment and a period of approximately 1360 years. In reality, maintaining a constant sedimentation rate over such time frames is unlikely. The rapid fluctuations in the benthic:planktonic ratio, evident in the diatom data of core LST1 for example, certainly suggests that dramatic and significant in-lake changes occurred during this time period which are likely to affect the sedimentation rate. Despite this though, the supposition that the gap between the cores is only slight is supported by the diatom data and the similar sedimentation rates of the cores.

In terms of interpreting a joint record from the lake, and based on the evidence presented, a gap of 3 cm between the cores has been assumed (this being the approximate mid-point between the 3.97 cm gap inferred by LSFS calculations and the 2.77 cm gap inferred by LST1 calculations). Based on this, the full Lake Surprise record presented here equates to a 344 cm core (with a 3 cm gap). The merged diatom record is presented in Figure 5.31, which illustrates dominant species that attained a relative abundance of 25% or more in any one sample, and Figure 5.32 which illustrates sub-dominant species which attained a relative abundance of 10 – 24% in an individual sample. The striking feature of the combined record is the shift in the state of the lake, from one of high variability (most evident in the record of benthic:planktonic ratios), to one indicating a more stable state with a lower amplitude of variability. This, in addition to the formulation of a chronology for the merged cores, is discussed in Section 6.1.2.

Figure 5.31: The fossil diatom record of the combined Lake Surprise cores showing species that attained a relative abundance of greater than 25% of the total count in any one sample. The grey line indicates the gap in the record.

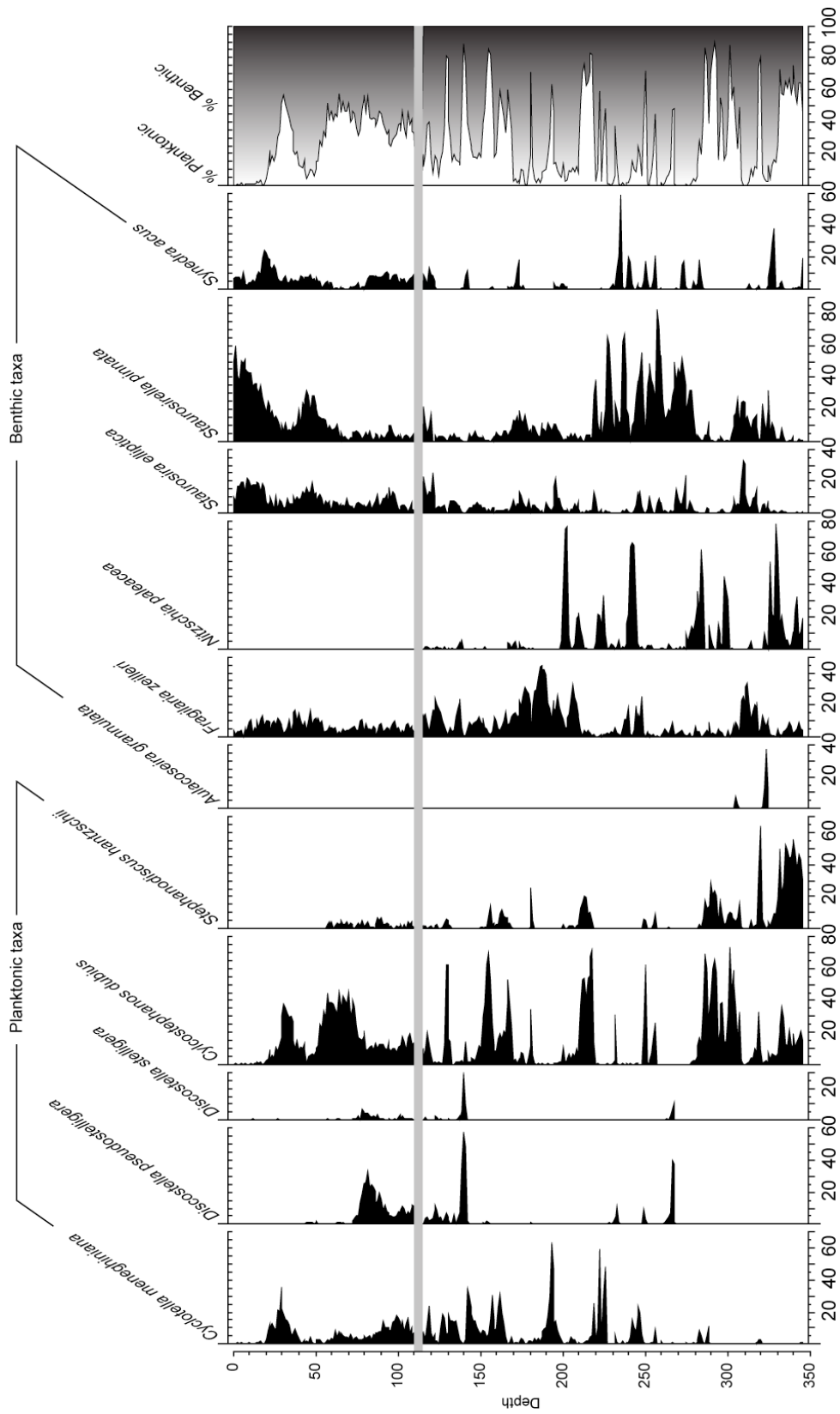
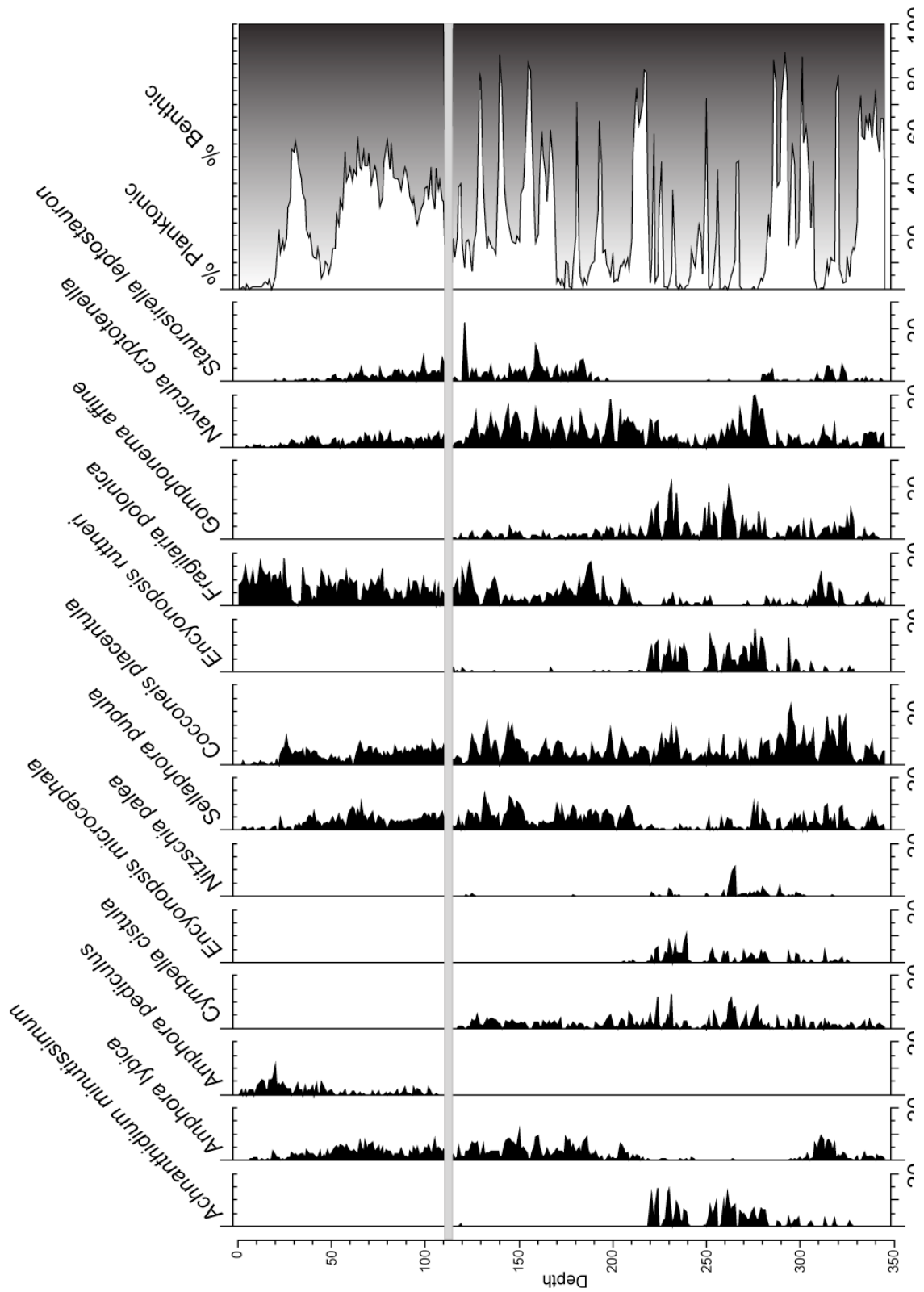


Figure 5.32: The fossil diatom record of the combined Lake Surprise cores showing all species that attained a relative abundance between 10% and 25% of the total count in any one sample. The grey line indicates the gap in the record.



## 5.4 Transfer function development

The development of the conductivity transfer function consisted of several steps. The first was to identify the minimum number of diatom frustules to count from each modern calibration sample to generate a robust model. This step is not usually reported in literature pertaining to transfer function development, though was undertaken here, firstly to test the common belief that counting more valves is better and, secondly, to maximise the potential of developing a robust transfer function particularly as the calibration set only contained a small number of sites ( $n = 47$ ). Following this, indirect gradient analysis determined if linear or unimodal methods of data analysis were most appropriate, with the outcome determining the next step in the process; which method of direct gradient analysis should be employed. Gradient analysis was employed to identify the environmental variable exerting the greatest influence on the diatom assemblages and a transfer function was developed for that dominant variable. The predictive capacity of, and size of errors in, the model may be improved through the removal of rare species and sites identified as outliers. These issues are examined before the final model is applied to the fossil diatom data to generate a reconstruction of conductivity. Methods of evaluating and validating the resulting reconstruction are also discussed.

### *5.4.1 Comparing count sizes*

The number of diatom valves enumerated in the development of modern calibration sets varies from study to study. However, assessing the value of increasing count sizes is rarely undertaken. As a means of assessing increasing count sizes in this study, cumulative totals of 100, 200, 300 and 400 valves were counted from the modern samples, thus effectively providing four modern datasets. In order to test the reconstructive potential of these datasets, a simple diatom-conductivity transfer function was developed using WA (classical and inverse deshrinking) and WA-PLS regression methods, with no rare species deleted. The jack-knifed  $r^2$  and RMSEP values for the four datasets were compared to identify the dataset with the best potential for further development.

The results, presented in Figure 5.33, illustrate that the highest jack-knifed  $r^2$  values (0.63) were achieved using simple WA techniques for counts of 200 and 300 valves, while the lowest RMSEP was clearly achieved using WA (classical) regression with

counts of 200 valves. In all cases, the WA-PLS model performed worse than the WA models in terms of the jack-knifed  $r^2$  correlation coefficient. Based on the results of this simple test, the 200-count dataset was identified as that which had the most potential for further development as it demonstrated a combination of the highest jack-knifed  $r^2$  with the lowest RMSEP. All further analyses were undertaken using this dataset.

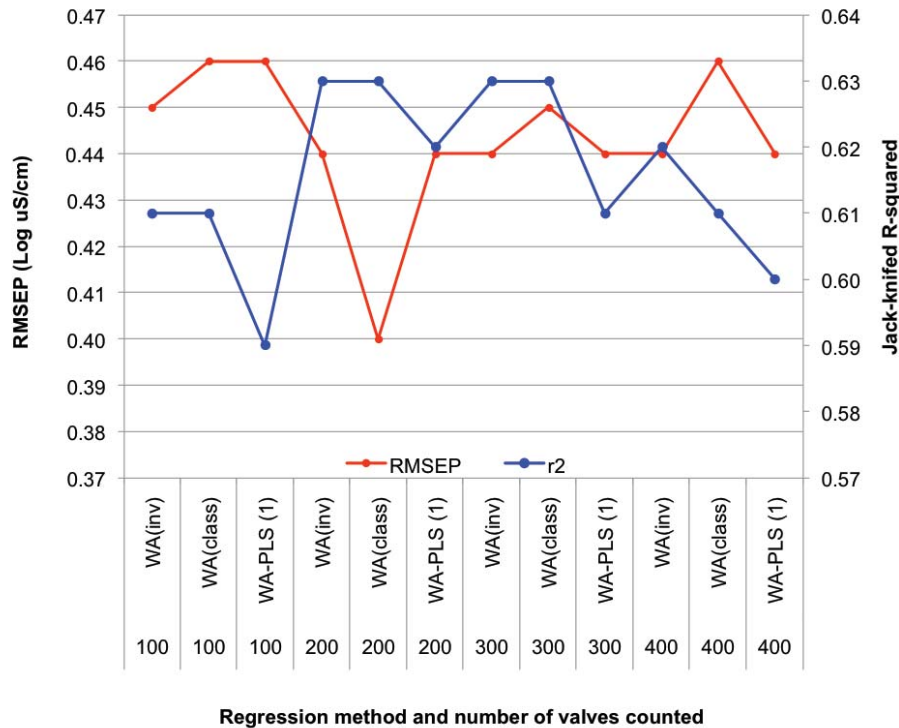


Figure 5.33: Results of increasing count size on jack-knifed  $r^2$  and RMSEP values.

In terms of enumerating modern samples for calibration sets, there seems to be little agreement on what constitutes a suitable number of valves to count. For example, Siver (1999), Kauppila *et al.* (2002) and Reid (2005) counted a minimum of 300 valves per sample, Ng and King (1999) and Davies *et al.* (2002) counted a minimum of 400 valves, Philibert and Prairie (2002), Yang *et al.* (2003) and Tibby (2004) counted a minimum of 500 valves, while Schönfelder *et al.* (2002) counted between 800 and 1300 valves per sample. A rare example of assessing the value of increasing count sizes is provide by Reavie and Smol (2001) who, prior to developing their transfer function, undertook a “preliminary rarefaction analysis” which determined that counting 300 valves would be sufficient for their study. This acknowledges that counting thresholds will be study-specific.

#### 5.4.2 Environmental characteristics of calibration set lakes

As outlined in Chapter 4, raw calibration set data was prepared for analysis through a number of steps, including the deletion of several sites for various reasons. The final dataset used for calibration consisted of a total of 47 sites with nine measured environmental variables. Table 5.14 provides the descriptive statistics for each variable across the calibration set while Table 5.15 shows the measured variables for each site .

*Table 5.14: Descriptive statistics for measured environmental variables in the calibration set*

<b>Variable</b>	<b>Minimum</b>	<b>Maximum</b>	<b>Median</b>	<b>Mean</b>	<b>Standard Deviation</b>
<b>Alkalinity</b> ( <i>log mg/L</i> )	0.00	3.25	2.30	2.21	0.56
<b>Calcium</b> ( <i>log+1 mg/L</i> )	0.02	3.15	1.40	1.50	0.58
<b>Conductivity</b> ( <i>log <math>\mu</math>S/cm</i> )	1.91	4.33	3.40	3.29	0.71
<b>Depth</b> ( <i>log cm</i> )	0.30	4.30	1.58	1.61	0.72
<b>Magnesium</b> ( <i>log mg/L</i> )	0.41	2.97	1.77	1.70	0.68
<b>Nitrate</b> ( <i>mg/L</i> )	1.00	39.00	4.40	5.53	5.71
<b>Phosphorus</b> ( <i>log+1 mg/L</i> )	0.00	0.42	0.07	0.11	0.11
<b>pH</b>	3.35	11.00	8.40	8.07	1.58
<b>Temperature</b> ( $^{\circ}$ C)	15.10	30.20	20.60	21.53	3.53

As mentioned previously, the primary aim in developing this transfer function was to provide more robust palaeosalinity reconstructions from fresh-mesosaline (0-20 g/L salinity, *sensu* Gasse *et al.*, 1987) systems than what was possible from an existing diatom-salinity transfer function from the same region (Gell, 1997). Therefore, sample sites were preferentially chosen based on their known conductivity, with the aim of achieving an even distribution of sites across a salinity gradient. Due to the demise of many lakes as a result of the drought, a more opportunistic approach was adopted and sites were selected to generate a frequency distribution along the salinity gradient that was skewed towards the low end. Similarly skewed distributions were noted for alkalinity, calcium, depth, magnesium and phosphorus. Skewed data can result in an arch-effect in a CCA (Legendre and Legendre, 1998), which, as mentioned in Chapter 2, can lead to the spacing of samples along the first axis being unrelated to the amount of change present, while also exaggerating the length of the second axis. Therefore,

with the exception of calcium and phosphorus, data for these variables were log-transformed in order to normalise their distribution. Calcium and phosphorus data sets were log+1 transformed (see Section 4.3.1). Frequency histograms for all measured variables are illustrated in Figure 5.34.

As can be seen from the presented frequency histograms, distribution across environmental gradients are generally unimodal in appearance, with varying degrees of skewness. The distribution of *logP+1* and *LogCond* are exceptions, with both variables exhibiting skewed distribution. However, these variables are less skewed post-transformation than in untransformed format. Transforming raw phosphorous data improved the skewness from 1.748 to 1.233 while conductivity distribution improved from 1.438 to -.284. The intentionally short conductivity gradient of 1.91 – 4.33 log  $\mu\text{S}/\text{cm}$  (81.3 – 21380  $\mu\text{S}/\text{cm}$ ) is comparable to several other salinity transfer functions from freshwater systems (e.g., Ng and King, 1999; Ryves *et al.*, 2002; Tibby and Reid, 2004).

Table 5.15: Measured water quality characteristics of all the modern sample sites in the final calibration set. Site names are provided in Tables 4.2 and 4.3.

Site code	Cond (log $\mu\text{S/cm}$ )	Alk (log mg/L)	Cal +1 (log mg/L)	Depth (log cm)	Mg (log mg/L)	N (mg/L)	P +1 (log mg/L)	pH	Temp $^{\circ}\text{C}$
BLk	2.80	2.26	1.65	4.30	1.31	3.50	0.00	8.31	16.7
1a	2.86	2.26	1.22	2.30	0.96	5.57	0.36	7.80	19.3
2a	3.34	2.26	1.81	2.08	1.62	5.71	0.08	7.04	19.7
3a	4.04	2.63	1.85	1.30	1.83	7.71	0.18	8.49	18.0
4a	4.20	2.66	1.33	1.48	2.11	6.61	0.11	10.2	18.6
6a	3.87	3.25	1.37	1.70	1.57	10.98	0.42	8.87	19.8
7a	4.15	2.84	1.30	1.70	2.54	4.69	0.39	8.79	21.8
8a	3.40	1.10	1.78	1.60	1.77	7.45	0.05	7.83	19.8
9a	4.03	2.50	2.05	2.15	2.58	3.86	0.02	8.49	19.3
10a	3.95	2.57	1.42	1.78	2.40	8.57	0.25	9.13	19.5
11a	4.16	2.85	1.78	2.36	2.46	5.97	0.08	8.78	19.3
12a	3.58	2.55	1.80	2.11	2.10	2.72	0.21	8.66	18.3
13a	3.89	2.60	1.36	2.40	2.59	3.14	0.02	9.58	21.8
14a	2.88	2.44	2.12	2.00	2.54	4.85	0.20	7.84	22.5
15a	4.16	2.87	1.28	2.18	2.38	2.50	0.02	7.79	19.5
16a	2.90	2.26	1.36	2.18	1.51	1.17	0.05	8.54	22.6
17a	3.54	2.41	1.84	2.53	2.14	3.92	0.02	8.01	21.4
18a	3.01	1.61	1.62	2.26	1.67	2.08	0.16	7.91	28.8
19a	2.34	2.32	1.38	3.08	1.37	1.41	0.02	8.75	23.4
34A/1	3.29	2.45	1.20	1.30	0.72	9.40	0.12	11.0	23.3
34A/2	3.27	2.48	1.30	0.90	1.63	12.00	0.15	9.63	26.3
56/4	2.46	1.65	1.48	0.30	1.74	6.90	0.22	5.30	17.6
93/1	2.01	1.60	0.02	0.30	0.48	2.50	0.07	5.55	20.7
93/2	1.91	1.54	0.79	0.30	0.70	3.00	0.22	5.60	20.2
106/1	2.77	2.22	1.11	1.11	1.08	4.40	0.26	8.96	25.0
106/2	2.78	2.28	2.08	1.23	2.08	4.40	0.07	8.06	26.1
106/4	2.76	2.20	1.15	1.58	1.08	4.60	0.30	7.22	24.1
107/4	3.75	2.16	2.18	1.52	2.20	5.80	0.06	9.17	21.2
107/5	3.74	1.60	2.08	1.53	2.15	4.20	0.08	9.27	19.8
115/1	2.19	1.70	0.76	1.78	0.79	1.90	0.06	5.15	18.3
115/3	2.16	1.54	0.76	1.78	0.76	1.70	0.04	5.46	15.6
119/2	3.68	2.30	2.21	1.52	2.04	5.50	0.05	8.40	20.1
119/3	3.68	2.34	2.21	1.63	2.04	4.00	0.01	8.11	20.6
136/2	2.27	1.90	1.00	1.34	0.99	4.50	0.13	5.63	22.4
136B/2	3.49	0.00	2.32	0.54	2.18	39.00	0.02	3.35	15.1
137/3	2.22	1.90	0.77	1.48	0.62	2.70	0.03	7.52	27.5
137/4	2.16	1.40	0.58	1.70	0.41	1.80	0.04	8.38	24.9
195/1	3.50	2.62	1.75	0.70	1.93	3.80	0.08	9.23	19.0
1002/2	4.32	2.00	3.15	0.64	2.97	14.00	0.03	8.50	28.0
3094/3	4.29	2.04	1.72	1.76	2.58	4.80	0.03	9.89	24.9
3094/4	4.33	2.58	1.60	1.62	2.62	4.70	0.00	9.70	24.7
7000/2	3.69	2.91	1.15	1.54	1.41	6.20	0.26	9.08	25.1
7011/2	3.16	2.41	1.04	1.40	1.23	4.10	0.05	9.73	25.7
7011/3	3.18	2.41	1.04	1.26	1.23	4.40	0.12	10.4	30.2
9003/3	2.69	1.95	0.58	1.20	0.57	4.70	0.02	5.88	19.7
9053/1	3.85	2.62	2.23	1.04	2.04	1.30	0.02	7.10	17.9
9053/2	3.79	2.56	2.05	1.18	2.08	1.00	0.00	7.18	17.9



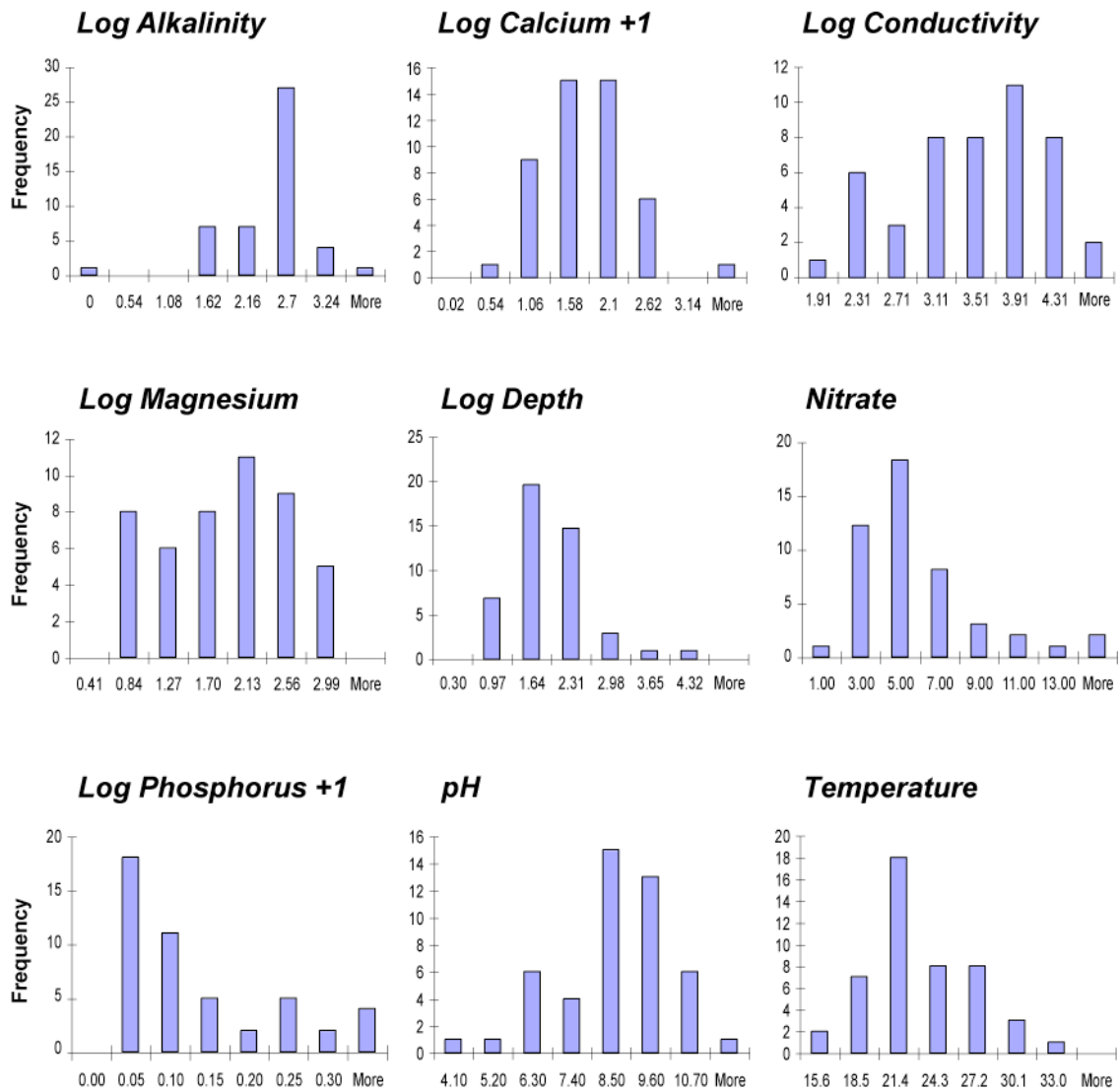


Figure 5.34: Frequency distributions of the measured environmental variables in the modern calibration set. Measurement units as per Table 5.15

Table 5.16 is a Pearson's correlation matrix of all environmental variables and indicates statistically significant correlations ( $p \leq 0.005$ ,  $r = \pm 0.403$ ) exist between a number of variables. Conductivity correlates most strongly with magnesium and calcium though also has significant correlations with magnesium and pH. Ideally, the variable of interest would have little or no covariance with the other measured variables, though multiple covariance between variables in calibration sets is common (Birks, 1998). For example, in their development of a chlorophyll-*a* transfer function, Jones and Juggins (1995) found that all nutrient variables were significantly correlated. Similarly, the dataset of Gell (1997), from the same region as this calibration set, demonstrated

significant correlations between salinity and all but two of the other 11 measured variables (145 samples,  $p \leq .005$ ,  $r \geq \pm .232$ ).

Table 5.16: Pearson's correlation matrix of environmental variables in the calibration set. Emboldened figures indicate a significant correlation at  $p \leq 0.005$ .

<b>log Alk</b>	1								
<b>log Cal</b>	0.125	1							
<b>log Cond</b>	<b>0.485</b>	<b>0.629</b>	1						
<b>log Depth</b>	0.297	0.134	0.091	1					
<b>log Mg</b>	0.309	<b>0.762</b>	<b>0.831</b>	0.118	1				
<b>Nitrate</b>	<b>-0.431</b>	0.309	0.225	-0.319	0.211	1			
<b>log+1 P</b>	0.278	-0.159	-0.033	-0.111	-0.115	0.054	1		
<b>pH</b>	<b>0.636</b>	0.227	<b>0.532</b>	0.280	0.324	-0.240	0.081	1	
<b>Temp</b>	0.115	-0.077	-0.074	-0.080	-0.081	-0.159	0.052	<b>0.459</b>	1
	<i>logAlk</i>	<i>logCal</i>	<i>logCond</i>	<i>logDepth</i>	<i>logMg</i>	<i>Nit</i>	<i>logP+1</i>	<i>pH</i>	<i>Temp</i>

#### 5.4.3 Analysis of diatom data: composition of surface sediment samples

In total, 169 taxa were recorded from the 47 sites which constituted the final modern calibration set. The average number of species per site was 15.6 (SD 6.4) with the maximum being 25 (sites 16a and 17a) and the minimum being just one (site 4a). Table 5.17 illustrates the number of species identified from each site. The average of 15.6 species/site is comparable to the study of Gell (1995) which had an average of 14.7 species per sample ( $n = 156$  samples) within the same geographic region, indicating that the lakes of the region generally display low diatom diversity. This is also apparent when considering that increasing the count size to 400 frustules (as described in Section 5.4.1 above) only added a further seven species from the 47 sites.

Table 5.17: Number of taxa recorded in surface sediment samples

Site Code	<i>n</i>	Site Code	No. of taxa	Site Code	No. of taxa
<b>BLk</b>	11	<b>17a</b>	25	<b>119/3</b>	18
<b>1a</b>	19	<b>18a</b>	19	<b>136/2</b>	15
<b>2a</b>	24	<b>19a</b>	22	<b>136B/2</b>	7
<b>3a</b>	9	<b>34A/1</b>	18	<b>137/3</b>	22
<b>4a</b>	1	<b>34A/2</b>	18	<b>137/4</b>	17
<b>6a</b>	13	<b>56/4</b>	10	<b>195/1</b>	4
<b>7a</b>	11	<b>93/1</b>	22	<b>1002/2</b>	11
<b>8a</b>	24	<b>93/2</b>	22	<b>3094/3</b>	6
<b>9a</b>	11	<b>106/1</b>	21	<b>3094/4</b>	5
<b>10a</b>	16	<b>106/2</b>	18	<b>7000/2</b>	7
<b>11a</b>	2	<b>106/4</b>	17	<b>7011/2</b>	19
<b>12a</b>	17	<b>107/4</b>	24	<b>7011/3</b>	14
<b>13a</b>	7	<b>107/5</b>	23	<b>9003/3</b>	20
<b>14a</b>	17	<b>115/1</b>	18	<b>9053/1</b>	16
<b>15a</b>	17	<b>115/3</b>	19	<b>9053/2</b>	18
<b>16a</b>	25	<b>119/2</b>	13		

As illustrated in Figure 5.35, sites with lower measured conductivity generally displayed a greater diversity of taxa. While the relationship is not strong, it is statistically significant ( $r^2 = 0.37$ ,  $p < 0.05$ ). Figure 5.36 demonstrates the response of the main species in the data set to increasing conductivity and illustrates species turnover along the gradient.

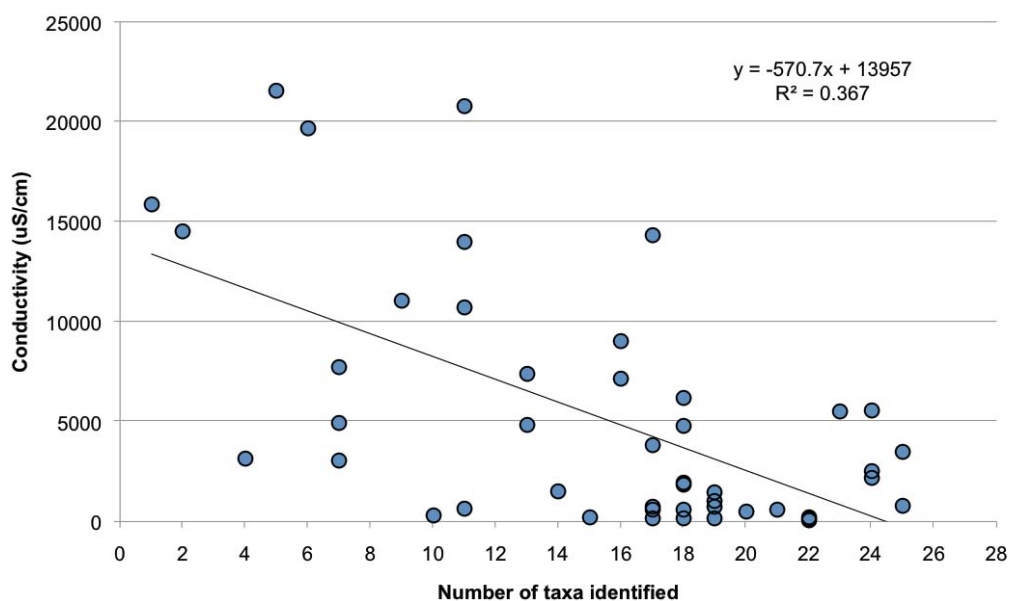


Figure 5.35: The relationship between measured conductivity and species diversity



Thirteen species occur in more than a quarter of all sites. The most commonly identified taxa were *Amphora veneta* and *Cocconeis placentula*, both of which were present in 23 sites. *Amphora veneta* attained greater than 1% relative abundance in 20 of the 23 sites, compared to 16 for *C. placentula*, though the former had a maximum relative abundance of only 18% (site 106/1), compared to 50.5% (site 3094/3) for the latter.

Sites 4a and 11a had the lowest taxonomic diversity with only one and two species identified respectively. Site 4a was the fourth most saline site (15,870  $\mu\text{S}/\text{cm}$ ) and registered the third highest pH (10.2). While it is, therefore, unsurprising that the diatom count included *Navicella pusilla*, a taxon indicative of brackish, alkaline environments (Gell *et al.*, 2002), it is somewhat surprising that it completely dominated the species assemblage. Interestingly, even when the count size was 400 valves, no other species were encountered. *Navicella pusilla* also attained a high relative abundance in sites 3094/4 (54% of the count), 1002/2 (53.5%) and 3094/3 (37.5%) which were the three most saline sites in the calibration set and were all alkaline (pH 9.7, 8.5 and 9.89 respectively).

The species identified in site 11a were *Cocconeis placentula* and *Entomoneis alata* which both registered 50% of the count. As mentioned above, *C. placentula* is one of the most commonly identified species in this dataset, and feature commonly in salinity transfer functions from a variety of locations (e.g., Gasse *et al.*, 1995; Gell, 1997; Reed, 1998; Ryves *et al.*, 2002), indicative of a broad salinity tolerance for this taxon. *Entomoneis alata* has also been shown to have a wide tolerance of salinity (Gell, 1997), though is only found in minor levels of abundance in other sites, accounting for only 3.5% and 3% in sites 3094/3 and 3a respectively. An outline of the autecology of the main species, with a focus on conductivity, is presented in Section 5.4.7.

#### 5.4.4 Indirect gradient analysis

Indirect gradient analysis, in the form of detrended correspondence analysis (DCA), was performed using CANOCO 4.5 (ter Braak and Šmilauer, 2002) to identify the manner in which diatom species were responding to changes in environmental variables. A DCA was performed on the full dataset of 169 species, with rare species downweighted. Results are presented in Table 5.18 and indicate that axes 1 and 2 account for 11.6% of the variance in species data. While this result is low, it is not unexpected given the large

number of taxa and the very large number of zero values (90% of all values). An axis 1 gradient length greater than 4 standard deviations indicates that a unimodal, rather than linear, method is most appropriate for direct gradient analysis (ter Braak and Prentice, 1988).

*Table 5.18: Results of DCA of species data from 47 sites.*

<b>Axes</b>	<b>1</b>	<b>2</b>	<b>3</b>	<b>4</b>	<b>Total Inertia</b>
<b>Eigenvalues</b>	0.896	0.802	0.648	0.450	<b>14.645</b>
<b>Lengths of gradient</b>	10.690	6.119	4.412	4.064	
<b>Cumulative % variance of spp. data</b>	6.1	11.6	16.0	19.1	
<b>Sum of all eigenvalues</b>					

#### 5.4.5 Direct gradient analysis

As established, a unimodal method of direct gradient analysis is most appropriate for this dataset. The most widely used unimodal direct gradient analysis technique (Palmer, 1993), and that employed in this project, is canonical correspondence analysis (CCA) which explores the relationship between the diatom taxa and the measured environmental variables. The process of CCA was undertaken on the full dataset to determine which environmental variable exerts the greatest influence over diatom species variation. The results informed the next step of the process, developing a transfer function for the dominant variable.

When the first CCA was performed, site 136B/2 was identified as a prominent outlier and was removed from further analyses. In all CCAs performed, automatic forward selection was enabled. This method of forward selection includes all the variables, in order of their influence on the species data, having been tested by 9999 Monte Carlo permutations. Results of the automatic forward selection are presented in Tables 5.19 and 5.20 and indicate that conductivity is the environmental variable exerting the greatest influence on diatom species variance. The  $\lambda_1$  result of 0.71 equates to 4.85% of species variance, which is a low result.

Table 5.19: Forward selection summary of CCA on full dataset with site 136B/2 deleted.

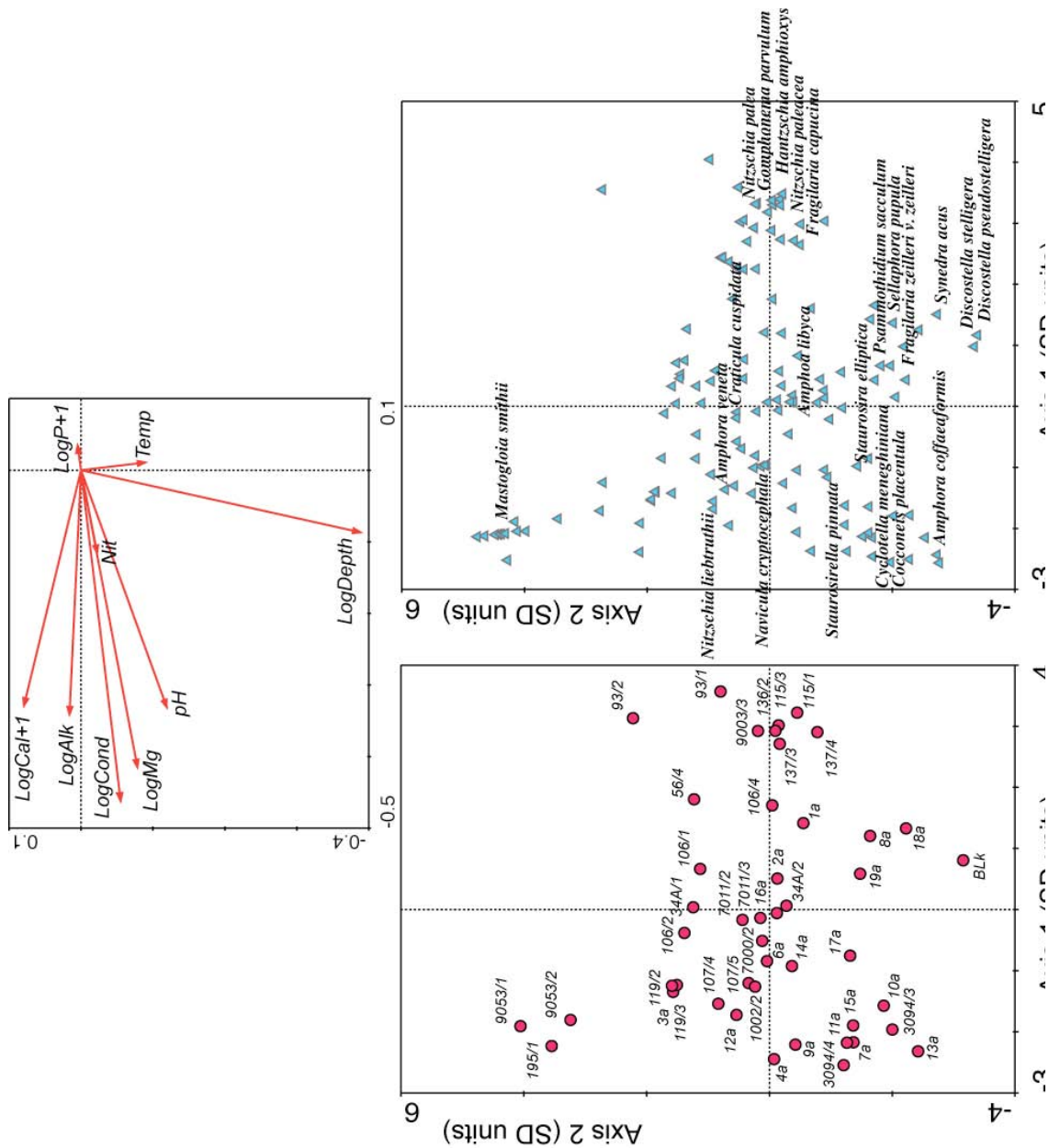
<i>Marginal Effects</i>					
<i>Variable</i>	<i>Var. No.</i>	<i>Lambda1</i>	<i>% variance explained</i>		
<i>LogCond</i>	3	0.71	4.85		
<i>LogMg</i>	5	0.67	4.57		
<i>pH</i>	8	0.58	3.96		
<i>LogDepth</i>	4	0.58	3.96		
<i>LogAlk</i>	1	0.57	3.89		
<i>LogCal+1</i>	2	0.53	3.62		
<i>Nit</i>	6	0.45	3.07		
<i>LogP+1</i>	7	0.40	2.73		
<i>Temp</i>	9	0.2	1.37		
<i>Conditional Effects</i>					
<i>Variable</i>	<i>Var. No.</i>	<i>LambdaA</i>	<i>p</i>	<i>F</i>	<i>% variance explained</i>
<i>LogCond</i>	3	0.71	0	2.35	4.85
<i>LogDepth</i>	4	0.57	0.001	1.91	3.89
<i>Nit</i>	6	0.43	0.023	1.46	2.94
<i>LogCal+1</i>	2	0.39	0.034	1.37	2.66
<i>pH</i>	8	0.39	0.046	1.34	2.66
<i>LogMg</i>	5	0.38	0.076	1.31	2.59
<i>LogAlk</i>	1	0.36	0.111	1.29	2.46
<i>LogP+1</i>	7	0.30	0.371	1.06	2.05
<i>Temp</i>	9	0.23	0.834	0.78	1.57

Table 5.20: Results of CCA of full dataset with site 136B/2 deleted.

<b>Axes</b>	<b>1</b>	<b>2</b>	<b>3</b>	<b>4</b>
<b>Eigenvalues</b>	0.778	0.663	0.543	0.465
<b>Species-environment correlations</b>	0.962	0.945	0.893	0.844
<b>Cumulative % variance of spp. data</b>	5.5	10.3	14.1	17.5
<b>Cumulative % variance of species-environment relation</b>	20.7	38.3	52.8	65.2
<b>Sum of all unconstrained eigenvalues</b>	<b>14.645</b>			
<b>Sum of all canonical eigenvalues</b>	<b>3.758</b>			

A graphical representation of the full CCA is presented in Figure 5.37, with the main modern and fossil species named.

Figure 5.37: CCA of full dataset with sample 136B/2 deleted



While further analysis of the data is possible, direct gradient analysis was performed primarily to identify the environmental variable exerting the most influence over diatom species' distribution. With this variable being identified as conductivity, a transfer function was then developed to enable conductivity reconstructions to be undertaken.



#### 5.4.6 Determining the best model for conductivity reconstructions

Prior to reconstructing conductivity, the model with the best reconstructive potential needs to be identified. Models can often be improved by removing the ‘noise’ of rare taxa or the extreme influence of outlying samples (Birks, 1998). Previously, it had been identified that counts of 200 frustules demonstrated the most potential for further development (see Section 5.4.1 above) and it was, by using this count number, that a number of species datasets were developed, with each dataset representing varying degrees of screening for rare taxa.

The criteria for identifying taxa as ‘rare’ varies from study to study and is usually determined in an *ad hoc* manner (Wilson *et al.* 1996). For example, Yang *et al.* (2003) exclude species that have less than 1% maximum abundance or only occur in one sample, Davies *et al.* (2002) identify rare species as being only those that occur in more than two samples and attain a maximum of 2% relative abundance, whereas Ryves *et al.* (2002) disregarded species that occurred in less than eight sites and less than 1% maximum abundance. In assessing the influence of rare species in this training set, five different species datasets were developed. The first (named ‘Full’) included all species regardless of their abundance, the second (‘>1%’) included only species that had attained a maximum abundance of greater than 1% in any sample, the third (‘>2%’) included only species that had attained a maximum abundance of greater than 2% in any sample, the fourth (‘>5%’) included species that had attained a maximum abundance of greater than 5% in any sample. The final dataset (‘1&2’) included only species that had occurred in more than one sample and had also attained a maximum abundance of greater than 2%.

Wilson *et al.* (1996:1585) state that “[t]he choice of various cut-off criteria [when identifying rare species] may affect the predictive abilities of different models (e.g., WA vs WA<sub>(tol)</sub>)”, therefore, for each dataset variation, a model was developed using weighted averaging (WA) regression utilising both classical and inverse deshrinking methods and weighted averaging – partial least squares (WA-PLS) regression. In each instance, leave-one-out cross validation (jack-knifing) was employed. For each model, outliers were iteratively identified and removed before re-running the model and testing the results by generating an  $r^2$  and RMSEP value. Outliers were defined as samples that had a jack-knifed residual value  $> \pm 1$  standard deviation of the environmental variable

being modelled (Gasse *et al.*, 1995; Jones and Juggins, 1995) and were removed in decreasing order of their residuals.

When developing a transfer function, Birks (1998) suggests a ‘minimum adequate model’ approach should be adopted in which an improvement in the RMSEP of 5% is required before undertaking data manipulation. A similar method could be applied in this instance where no further outliers are removed if the improvement in the RMSEP is less than 5%. However, while an improvement of less than 5% (range 3.7% - 4.4%) was noted in all the models tested here when sample 19a was removed as an outlier, it was decided to continue removing outliers to ascertain if this lead to a substantial reduction in the RMSEP and/or an increase in the  $r^2$ . As is evident in Table 5.21, removing the next outlier following 19a markedly improved the model, with a reduction in the RMSEP occurring in all cases and an average improvement of 13.3% noted (maximum 15.9%, minimum 6.8%, SD 2.4%). A considerable improvement in the  $r^2$  value was also noted in all cases. Given the substantial improvement in the models, the process of outlier removal continued until no further outliers were present.

By considering both classical and inverse deshrinking WA models, and using only one Component in WA-PLS models (as this was consistently the best performing WA-PLS model), 15 potential models were developed, three from each of the five species datasets. For each dataset, the WA-PLS model returned the highest RMSEP and these were therefore discounted. In terms of  $r^2$  and RMSEP values, little separated the classical deshrinking and inverse deshrinking WA models. In this instance, classical deshrinking methods were given preference as the inverse deshrinking method introduces bias at the ends of the environmental gradient (ter Braak and Juggins, 1993) and the fossil sites are ‘located’ at the dilute end of the gradient. Therefore, only the five best performing models using WA and classical deshrinking methods were identified based on their RMSEP and  $r^2$  values. These are presented in Table 5.22, though it is apparent that very little separates them.



Table 5.22: The five best performing WA (classical) models based on RMSEP values.

Rank	Species dataset	Jack-knifed $r^2$	RMSEP (log $\mu\text{S/cm}$ )
1	>2%	0.89	0.238
2	>5%	0.89	0.239
3	Full	0.88	0.239
3	>1%	0.88	0.239
3	1&2	0.88	0.239

In order to identify which of these five models had the best reconstructive potential at the low end of the conductivity gradient, the mean of the cross-validated residuals that fell below the observed conductivity of Lake Elingamite (3.54 log  $\mu\text{S/cm}$ ) were calculated for each model (*sensu* Birks *et al.*, 1990; Birks, 1998). Lake Elingamite was chosen as it had the highest observed conductivity of the two fossil study sites. The models with the equal lowest mean residual were those from the ‘Full’ and ‘>1%’ datasets, though the model from the ‘>2%’ dataset returned a mean residual result only marginally greater (Table 5.23).

Table 5.23: The five best models ranked in order of their average residuals less than the observed conductivity of Lake Elingamite.

Model Number	Species Dataset	No. of sites	No. of species	Jack-knifed $r^2$	RMSEP (log $\mu\text{S/cm}$ )	Average of residuals less than 3.54 log $\mu\text{S/cm}$ (the observed conductivity of Lake Elingamite)
1	Full	40	151	0.88	0.239	.117
2	>1%	40	139	0.88	0.239	.117
3	>2%	40	118	0.89	0.238	.118
4	1&2	40	104	0.88	0.239	.124
5	>5%	40	89	0.89	0.239	.125

As models 1, 2 and 3 had equitable average residuals below the observed conductivity of Lake Elingamite, the model that was chosen for all conductivity reconstructions performed in this project was Model 3 (>2%). This was primarily due to the slightly higher jack-knifed  $r^2$  and marginally lower RMSEP than models 1 and 2. Further details

of Model 3 are provided in Table 5.24 and demonstrate the impact of the jack-knifing technique on the  $r^2$  and errors.

Table 5.24: Details of the model identified as the most appropriate for conductivity reconstructions.

Name	RMSE (log $\mu\text{S/cm}$ )	Apparent $r^2$	Jack-knifed $r^2$	Jack-knifed Average Bias (log $\mu\text{S/cm}$ )	Jack-knifed Maximum Bias (log $\mu\text{S/cm}$ )	RMSEP (log $\mu\text{S/cm}$ )
Model 3	0.133	0.97	0.89	0.009	.319	0.238

Figure 5.38 illustrates both the apparent inferred conductivity, and the jack-knifed inferred conductivity, plotted against the observed conductivity (as log  $\mu\text{S/cm}$ ) for the 40 sites in the final model. The residuals for both the apparent and jack-knifed models are also presented with the jack-knifed residuals, displaying no obvious trends.

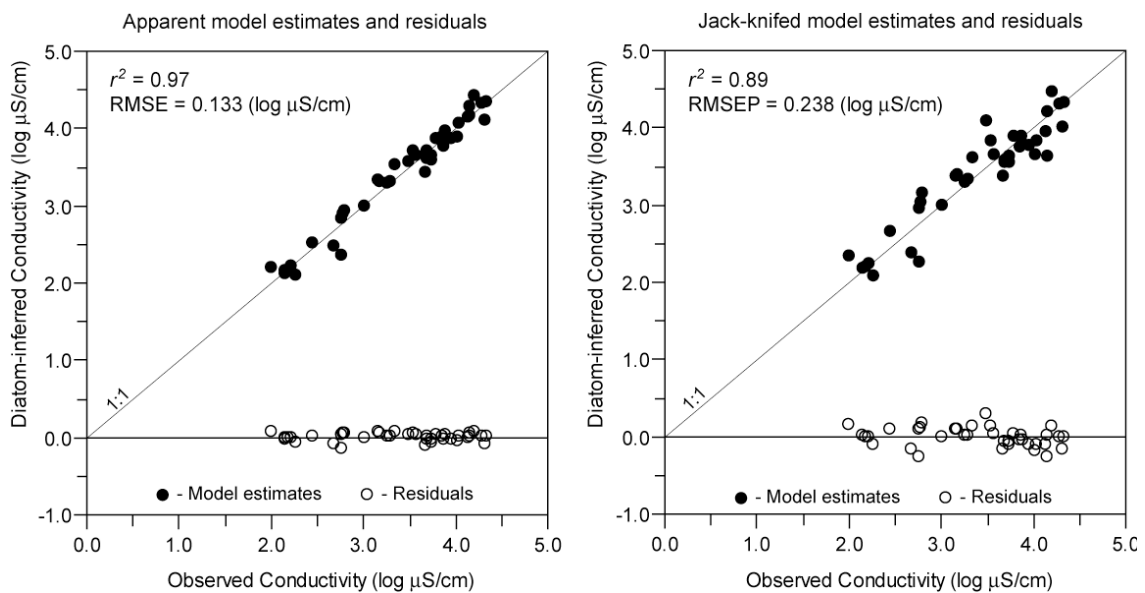


Figure 5.38: Diatom-inferred conductivity and residuals against observed conductivity from both the apparent and jack-knifed WA (classical) regression.

#### 5.4.7 Diatom taxon tolerances and optima

The diatom conductivity tolerance and optima are reported in log  $\mu\text{S}/\text{cm}$ . Values are back-transformed for comparison with other studies. Figure 5.39 illustrates the back-transformed conductivity tolerance and optima of the most abundant taxa. All but four of these species have a conductivity optima less than 10,000  $\mu\text{S}/\text{cm}$ .

The most common species in the calibration set are briefly discussed below in terms of their conductivity optima. Results are also contrasted against those from other studies as a means of assessing the transfer function. Thirteen species occurred in approximately a quarter ( $n = \geq 12$ ) or more of all sites and are presented in order of the frequency at which they occurred. Inter-study comparisons are confounded by the use of varying measures of conductivity/salinity. Therefore, to enable comparisons, measurements of conductivity have been converted to  $\mu\text{S}/\text{cm}$ , while measurement of salinity reported in g/L have been converted to  $\mu\text{S}/\text{cm}$  using the following equation;

$$(x \text{ g/L} \times 0.68) \times 1000$$

The factor of 0.68 follows that of Gell (1995). Given that conductivity is a measure of ionic composition, while salinity is a measure of the mass of ionic compounds per litre of water (Williams, 1986), it is recognised that there are limitations to applying such an all-inclusive conversion factor in order to compare two differing variables. The conversions noted below are therefore not absolute in their accuracy, however, for the purpose of aiding comparisons between species tolerances and optima, this conversion method is considered adequate.

#### *Amphora veneta*

Together with *Cocconeis placentula*, this was the most commonly encountered species, occurring in 23 of the 47 sites. Obtaining an optimum of 3.163 log  $\mu\text{S}/\text{cm}$  (1458  $\mu\text{S}/\text{cm}$ ), with a tolerance of  $\pm 0.497$  log  $\mu\text{S}/\text{cm}$ , it attained a maximum relative abundance of 18% in site 106/1. Gell (1995) calculated a salinity optimum of 1.95 g/L (approximately 1326  $\mu\text{S}/\text{cm}$ ) with similar results found by Wilson *et al.* (1996) (1.79 g/L;  $\sim 1217$   $\mu\text{S}/\text{cm}$ ), though Yang *et al.* (2003) identified a lower optimum of 598  $\mu\text{S}/\text{cm}$ . Reed (1998) has identified two forms of the species with optima ranging from 7150 to 7690  $\mu\text{S}/\text{cm}$ . *A. veneta* appears to prefer alkaline conditions, occurring in this calibration set at higher concentrations in waters with a pH >8. This is consistent

with the findings of Gell (1995) and Gasse *et al.* (1995) who calculated pH optima of 8.4 and 8.14 respectively.

#### *Cocconeis placentula*

As with *Amphora veneta*, this taxon occurred in 23 of the 47 sites in the calibration set, however it attained a greater maximum relative abundance of 50.5% (site 3094/3) and also occurred at 50% (site 11a), 37.5% (site 3094/4) and 31.6% (site 7a). Its derived optimum was 4.142 log  $\mu\text{S/cm}$  (13,882  $\mu\text{S/cm}$ ), though it had a wide range of tolerance from 6447  $\mu\text{S/cm}$  to 29,819  $\mu\text{S/cm}$ . Optima attained in other studies vary significantly. Yang *et al.* (2003) calculated an optimum of 765  $\mu\text{S/cm}$ , while Gasse *et al.* (1995) calculated an optimum of 467  $\mu\text{S/cm}$ . Wilson *et al.* (1996) identified the taxon to variety level and found optima ranged from 197  $\mu\text{S/cm}$  to 217  $\mu\text{S/cm}$ , while Reed (1998) obtained an optimum of 20,660  $\mu\text{S/cm}$  for *C. placentula* var. *euglypta*. Gell (1995) identified a weighted averaged optimum of 12.0 g/L ( $\sim 8,160 \mu\text{S/cm}$ ) at species level identification, though found this taxa to be abundant in waters of both less than 1 g/L and greater than 100 g/L (680 – 68,000  $\mu\text{S/cm}$ ).

#### *Cyclotella meneghiniana*

*Cyclotella meneghiniana* is a very commonly found diatom in lakes and rivers around the world (Shafik *et al.*, 1997). In this dataset it was found in 21 of the 47 sites, attaining a maximum relative abundance of 38.5% in site 10a, an alkaline, phosphorus and nitrogen rich site (Table 5.15). The conductivity optimum was 3.631 log  $\mu\text{S/cm}$  (4278  $\mu\text{S/cm}$ ), with a tolerance range from 1802 to 10,117  $\mu\text{S/cm}$ . This finding is within the boundaries of the optima identified in other studies, falling midway between the optima of Reed (1998; 8640  $\mu\text{S/cm}$ ) and Gasse *et al.* (1995; 6025  $\mu\text{S/cm}$ ) at one end and those of Gell (1995; 3.4 g/L ( $\sim 2312 \mu\text{S/cm}$ )), Wilson *et al.* (1996; 1.37 g/L ( $\sim 932 \mu\text{S/cm}$ )) and Tibby and Reid (2004; 407  $\mu\text{S/cm}$ ) at the other.

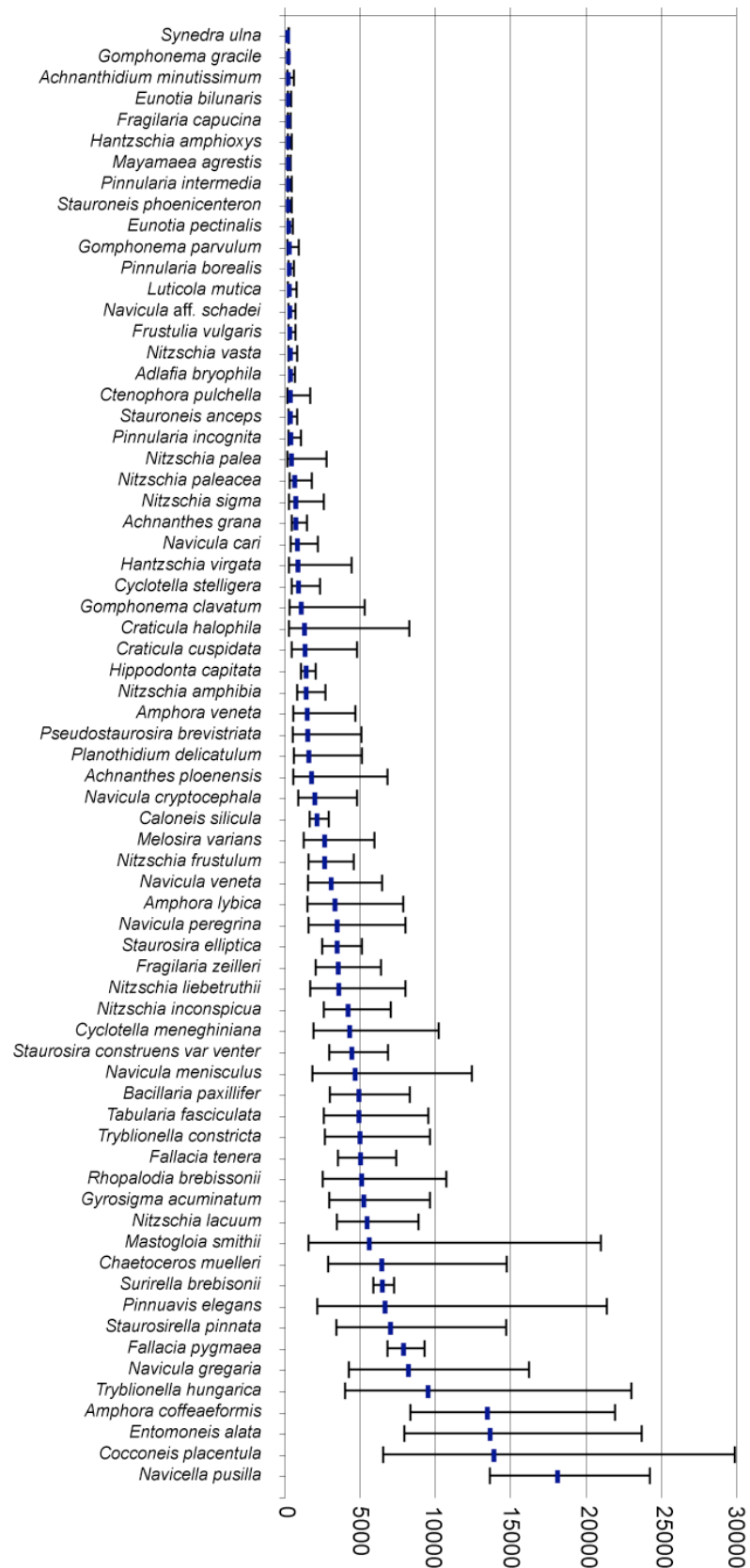


Figure 5.39: The back-transformed conductivity optima and tolerance to conductivity, derived by weighted averaging. Only taxa that attained a maximum relative abundance >3% and were present in 3 or more samples are presented (n = 69).



### *Nitzschia liebtruthii*

This taxon was present in 19 sites and, where it did occur, it was usually dominant with an average relative abundance of 11%. Its derived optimum was 3.549 log  $\mu\text{S}/\text{cm}$  (3541  $\mu\text{S}/\text{cm}$ ), with a tolerance of 1583 to 7928  $\mu\text{S}/\text{cm}$ . Gell (1995:202) describes this species as being “perhaps the most eurytopic” in his dataset. In terms of its salinity preference, this is reflected by the varying optima obtained by this and other studies. Gell (1995) obtained a WA optimum of 7.8 g/L ( $\sim$ 5304  $\mu\text{S}/\text{cm}$ ) and a tolerance of  $\pm$  3.9g/L ( $\sim$ 2652  $\mu\text{S}/\text{cm}$ ), Reed (1998) calculated 14,330  $\mu\text{S}/\text{cm}$  with a tolerance range of 33,600  $\mu\text{S}/\text{cm}$ , while Wilson *et al.* (1996) derived an optimum of 18.66 g/L ( $\sim$ 12,688  $\mu\text{S}/\text{cm}$ ) for the nominate variety of the species, though noted an enormous tolerance ranging from 0.65 to 535.8 g/L (442 – 364,344  $\mu\text{S}/\text{cm}$ ). Interestingly, the tolerance ranges increase in line with the length of the salinity gradient in these studies, suggesting that Gell (1995) is correct in classifying this as a eurytopic species.

### *Gomphonema parvulum*

This taxon was found in 18 sites, attaining a maximum relative abundance of 20.5% in site 93/1, a relatively fresh (102  $\mu\text{S}/\text{cm}$ ), acidic (pH 5.5) site. Interestingly, site 93/1 is also characterised by low nutrient values (N = 2.5 mg/l, P = 0.18 mg/l), though *G. parvulum* is known to be tolerant of high levels of organic pollution (Kelly, 1998), suggesting a large range of tolerance to nutrients. In terms of conductivity, this taxon commonly displays a preference for low conductivity waters. Optima range from 197  $\mu\text{S}/\text{cm}$  (0.29 g/L; Wilson *et al.* 1996) to 864  $\mu\text{S}/\text{cm}$  (Yang *et al.*, 2003) and 952  $\mu\text{S}/\text{cm}$  (1.4 g/L; Gell, 1995). The optimum derived here was 2.384 log  $\mu\text{S}/\text{cm}$  (242  $\mu\text{S}/\text{cm}$ ).

### *Craticula cuspidata*

While this taxon could be classed as ‘common’, being present in 17 sites, it was rarely dominant. Of the 17 sites, it only attained a relative abundance greater than 1% in seven sites, with a maximum of just 7% in site 34A/1, the most alkaline site in the dataset (pH 11). The derived conductivity optimum was 3.11 log  $\mu\text{S}/\text{cm}$  (1288  $\mu\text{S}/\text{cm}$ ), which is higher than the 516  $\mu\text{S}/\text{cm}$  (0.76 g/L) of Wilson *et al.* (1996) and the 977  $\mu\text{S}/\text{cm}$  of Gasse *et al.* (1995), though less than the 3.55 g/L (2412  $\mu\text{S}/\text{cm}$ ) of Gell (1997).

### *Nitzschia palea*

Present in 14 sites, the derived conductivity optimum for *N. palea* was 2.165 log  $\mu\text{S/cm}$  (412  $\mu\text{S/cm}$ ). Similarly fresh optima were identified Wilson *et al.* (1996; 0.6 g/L; ~ 408  $\mu\text{S/cm}$ ), Gasse *et al.* (1995; 891  $\mu\text{S/cm}$ ) and Ng and King (1999; 159  $\mu\text{S/cm}$ ).

### *Tabularia fasciculata*

The derived optimum for this taxon varies greatly within the literature. Gasse *et al.* (1995) found an optimum of 45,708  $\mu\text{S/cm}$ , while Wilson *et al.* (1996) derived an optimum of approximately 6072  $\mu\text{S/cm}$  (8.93 g/L). Reed (1996) identifies two unofficial types in her dataset, with widely differing conductivity optima of 11,380  $\mu\text{S/cm}$  and 47,480  $\mu\text{S/cm}$ . The derived optimum from this dataset was 3.688<sub>log</sub> $\mu\text{S/cm}$  (4873  $\mu\text{S/cm}$ ), slightly higher than the optimum of 3.31 g/L (2250  $\mu\text{S/cm}$ ) derived by Gell (1997).

### *Navicula cryptocephala*

Generally preferring fresh water environments less than 2,000  $\mu\text{S/cm}$  (Sonneman *et al.*, 2000), it was surprising to find this species dominating sites 7011/3 (50.2%) and 7011/2 (34.3%) which had conductivity values of 15,130  $\mu\text{S/cm}$  and 14,420  $\mu\text{S/cm}$  respectively. This may account for the derived conductivity optimum of 3.288 log  $\mu\text{S/cm}$  (1940  $\mu\text{S/cm}$ ) which, when compared to other studies, is high. These range from 0.09 g/L (61  $\mu\text{S/cm}$ ) of Wilson *et al.* (1996) to the 1227  $\mu\text{S/cm}$  of Ng and King (1999). Gasse *et al.* (1995) obtained a conductivity optimum of 407  $\mu\text{S/cm}$ , similar to that of Yang *et al.*'s (2003) 450  $\mu\text{S/cm}$  optimum. Interestingly, from the same study area as this dataset, Gell (1995) derived a WA optimum of 3.1 g/L, approximately 2108  $\mu\text{S/cm}$ , similar to the findings of this study.

### *Navicula menisculus*

Identified in 13 sites with a maximum relative abundance of 17% in site 7a, this taxon had a conductivity optimum of 3.665 log  $\mu\text{S/cm}$  (4623  $\mu\text{S/cm}$ ) and a tolerance range of 1726 to 12,346  $\mu\text{S/cm}$ . Comparisons with other studies are difficult as only Gasse *et al.* (1995) provide information for this species. They identify a conductivity optimum of 181  $\mu\text{S/cm}$ , considerably lower to that obtained in this study. In a guide to common diatoms from temperate Australian streams, Sonneman *et al.* (2000) suggest this taxon has a preference for waters with a conductivity greater than 2000  $\mu\text{S/cm}$ , supporting the findings of this study.

### *Fragilaria capucina*

This taxon was not identified in the modern dataset of the region by Gell (1995, 1997), though the European Diatom Database (EDDI; <http://craticula.ncl.ac.uk/eddi>) plots the salinity optima of the taxon at approximately  $2.5 \log_{10}$  Conductivity ( $\sim 315 \mu\text{S/cm}$ ). Yang *et al.* (2003) identified an optimum of  $390 \mu\text{S/cm}$  and a preference of water depth of less than one metre, while Ng and King (1999) identified an optimum of  $1227 \mu\text{S/cm}$ . In the case of the latter though, this calculation is made on the basis of only three occurrences in the dataset.

In the modern dataset presented here, this taxon occurred in 12 sites attaining a maximum abundance of 27.4% in site 137/4, though it was also dominant in sites 137/3 (24.5%) and 136/2 (25.5%). Respectively, these sites are the 4<sup>th</sup>, 6<sup>th</sup> and 7<sup>th</sup> freshest sites in the dataset. While sites 137/3 and 137/4 were alkaline with pH of 7.52 and 8.38 respectively, site 136/2 was acidic (pH 5.63), indicating that this taxon has a wide pH tolerance. The conductivity optimum for this taxon was  $2.25 \log \mu\text{S/cm}$  ( $178 \mu\text{S/cm}$ ) with a tolerance of  $\pm 0.22 \log \mu\text{S/cm}$ , comparable to results reported elsewhere. According to Kelly *et al.* (2005), this species has a preference for “circumneutral water with low to moderate nutrient levels.” This assertion that is supported by Tibby (2004) who found the species to be most abundant in mesotrophic environments.

### *Hantzschia amphioxys*

The maximum relative abundance attained by this taxon was 14.3% in site 115/1. It was present in 12 sites of the training set, though only attained greater than 1% abundance in six sites. Its derived optimum is  $2.262 \log \mu\text{S/cm}$  ( $182.7 \mu\text{S/cm}$ ), with a tolerance of  $\pm 0.286 \log \mu\text{S/cm}$ . Examination of other studies shows wide ranging outcomes. Gasse *et al.* (1995) report an optimum of  $478 \mu\text{S/cm}$ , while at the other end of the scale, Reed (1995) found an optimum of  $39,210 \mu\text{S/cm}$ . Wilson *et al.* (1996) derived an optimum of  $4.62 \text{ g/L}$  ( $\sim 3141 \mu\text{S/cm}$ ), though note a very wide range from  $0.15$  to  $139.64 \text{ g/L}$  ( $\sim 102 - 94,955 \mu\text{S/cm}$ ). The widely varying reported optima suggests that this taxon has a very broad tolerance to salinity.

### *Navicula veneta*

Present in 12 sites, *N. veneta* was most common in site 119/2, accounting for 14.1% of the total count. It appears to have a wide pH tolerance, occurring at >5% abundance in sites with pH values ranging from 5.6 to 11. Its derived conductivity optimum was  $3.483 \log \mu\text{S/cm}$  (3040  $\mu\text{S/cm}$ ), with a range of 1448 to 6372  $\mu\text{S/cm}$ . While this optimum is higher than the 2.4 g/L (~1632  $\mu\text{S/cm}$ ) of Gell (1997) and the 1842  $\mu\text{S/cm}$  of Wilson *et al.* (1996), it is below the optimum of 9350  $\mu\text{S/cm}$  derived by Reed (1998).

#### 5.4.8 Comparisons of model performance with other studies

As is evident from the discussion above, the derived conductivity optima of the main species identified generally compare well with other diatom-salinity transfer functions. However the  $r^2$  and RMSEP values, in conjunction with patterns of residual distribution, provide a further illustration of the strength of the model. As illustrated in Figure 5.38 above, the residuals for the preferred model presented here show no obvious trends of under- or over-estimation at either end of the gradient. Table 5.25 shows how the results of this study compare with other diatom salinity transfer functions in terms of  $r^2$  (apparent and jack-knifed), RMSE and RMSEP as a proportion of gradient length. Comparisons of both the apparent and jack-knifed  $r^2$  values show that this study performs very well, with only Yang *et al.* (2003) returning a higher value. Similarly, the measure of RMSEP as a proportion of gradient length also indicates that only the Yang *et al.* (2003) and Ryves *et al.* (2002) studies are better than the outcomes of this study.

Table 5.25: Comparisons between the diatom conductivity transfer function performance in this and other studies. Where studies reported errors as log data, gradient length was determined by subtracting the log-minimum value from the log-maximum value of the stated conductivity gradient. RMSE and RMSEP units are the same as those for Gradient length. n.r. – denotes ‘not reported’. \* - log value provided for ease of comparison

Study	Region	$r^2_{\text{apparent}}$	RMSE	$r^2_{\text{Jack}}$	RMSEP	Gradient length	RMSEP as % of gradient length
This study	Western Victoria, Australia	0.97	0.133	0.89	0.238	2.42 log $\mu\text{S}/\text{cm}$	9.8%
Gasse <i>et al.</i> (1995)	Africa	0.87	0.32	0.81	0.39	3.39 log $\mu\text{S}/\text{cm}$	11.5%
Wilson <i>et al.</i> (1996)	Western North America	n.r	n.r	0.87	0.371	n.r	
Gell (1995, 1997)	Western Victoria, Australia	0.79	0.2886	0.72	0.3357	2.43 log g/L	11.9%
Reed (1998)	Spain	0.91	0.18	0.57	0.41	3.35 log mS/cm	12.2%
Ryves <i>et al.</i> (2002)	Greenland	0.96	0.12	0.88	0.217	2.23 log $\mu\text{S}/\text{cm}$	9.7%
Yang <i>et al.</i> (2003)	Tibetan Plateau	n.r	n.r	0.92	0.22	2.99 log $\mu\text{S}/\text{cm}$	7.4%
Tibby & Reid (2004)	Murray River, Australia	n.r	n.r	0.71	114	1012 $\mu\text{S}/\text{cm}$ (3.01 log $\mu\text{S}/\text{cm}$ )*	11.2%
Reid (2005)	New Zealand	n.r	n.r	0.79	0.151	1.42 log $\mu\text{S}/\text{cm}$	10.6%

The evaluation of this transfer function indicates that it performs well, with a high jack-knifed  $r^2$  value and a low RMSEP. The transfer function can now be applied to fossil data for down-core reconstructions of conductivity.

## 5.5 Transfer function application

### 5.5.1 Conductivity reconstructions

Using the model described above, reconstructions of conductivity were derived for the three study site cores. Results are presented in Figure 5.40 (Lake Elingamite, Core LE1), Figure 5.41 (Lake Surprise, Core LSFS) and Figure 5.42 (Lake Surprise, Core LST1) with diatom-inferred conductivity measurements back-transformed from log data. Reconstruction diagnostics for cores LE1, LSFS and LST1 are presented in Appendices 8, 9 and 10 respectively.

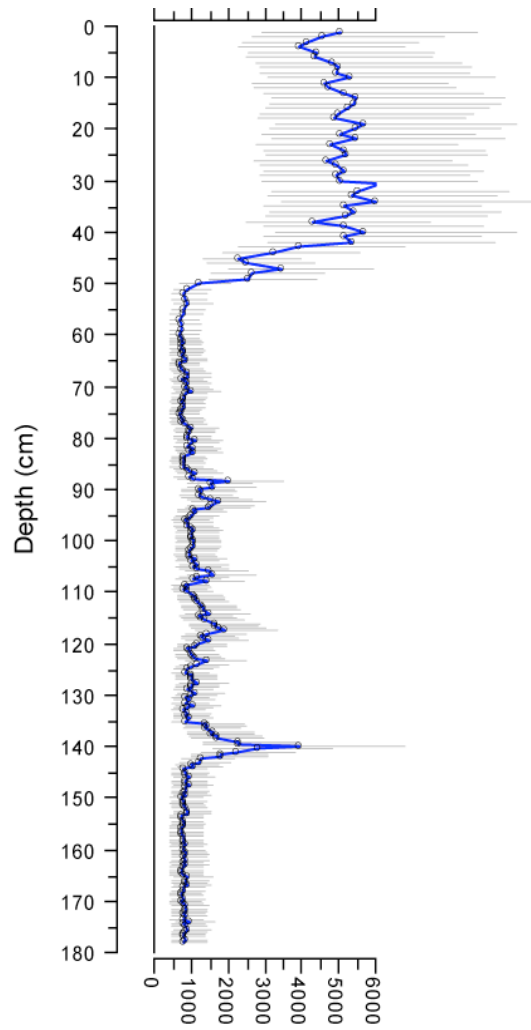


Figure 5.40: Diatom-inferred conductivity reconstruction from Lake Elingamite, core LE1.

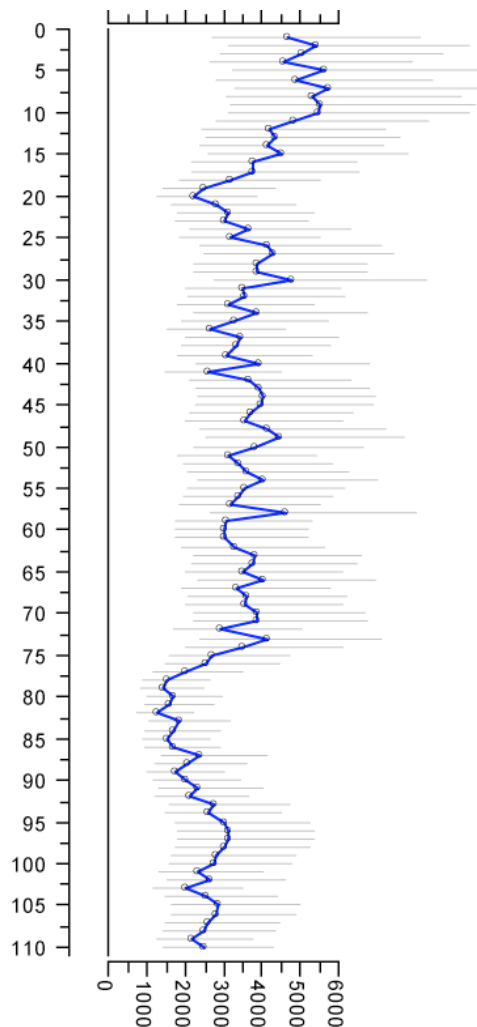
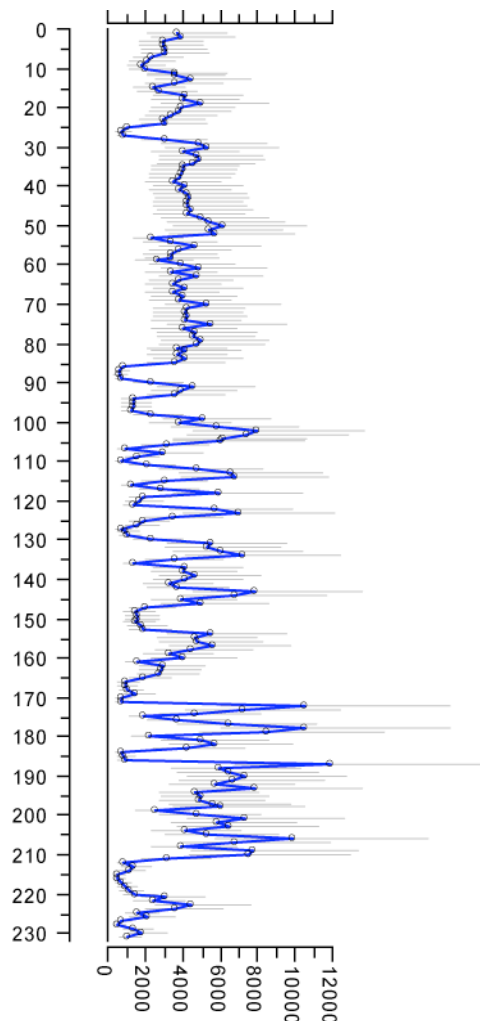


Figure 5.41: Diatom-inferred conductivity reconstruction from Lake Surprise, core LSFS

### 5.6 Evaluation and validation of conductivity reconstructions

In terms of evaluating diatom water quality models, Birks (1998:313) states that “the most powerful means of validation is to compare the reconstructions, at least for the recent past, against known recorded historical environmental records”. There are a number of examples where this has been undertaken successfully (e.g., Fritz, 1990; Laird *et al.*, 1996b; Laird *et al.*, 1998), though it is more often the case that such comparisons are not possible (Davies *et al.*, 2002). This can be due to a number of reasons, such as a lack of contemporary records, insufficient or sporadic monitoring, or a level of anthropogenic catchment disturbance (e.g., eutrophication, acidification, salinisation) that fundamentally alters the diatom community thereby obscuring much taxon-based evidence of other changes that may be of interest (e.g., climate). In the event where direct historical comparisons are not possible, the accuracy of

reconstructions can be evaluated through numerical method such as the  $r^2$  correlation coefficient and the RMSEP (Birks, 1998).



*Figure 5.42: Diatom-inferred conductivity reconstruction from Lake Surprise, core LST1*

While the  $r^2$  and RMSEP indicate the reconstructive strength of the overall model, individual reconstructed samples may still be poorly predicted due to having poor, or no, analogues in the calibration set. Logically, fossil samples with close analogues in the modern calibration set are likely to provide a more reliable reconstruction. Conversely, reconstructions from fossil samples with no, or poor, modern analogues can not be interpreted with confidence.

A number of methods exist to assess analogue quality. One simple method is the identification of the proportion of species in each fossil sample that are absent from the



calibration set, or are 'rare' or poorly represented in the calibration set (e.g., occur in less than 10% of samples) (Birks, 1998). This can also be expanded by determining sample-specific errors for fossil samples which will provide a more realistic measure of error than a 'whole-model' RMSEP (Ryves *et al.*, 2002). The 'goodness-of-fit' method is another measure that is undertaken by passively plotting the fossil samples in a CCA that is constrained to the single variable of interest and assessing the residual distance of the plotted samples to the axis as compared to that of the training set (e.g., Birks *et al.*, 1990; Laird *et al.*, 1996b).

The modern analogue technique (MAT) is a further measure of evaluating the reconstructions and involves measuring the dissimilarity coefficient of the distance between samples in the calibration set and comparing the results with the distance between individual fossil samples and those in the calibration set. Dissimilarity Coefficients (DCs) can be measured using either the  $x^2$  distance dissimilarity coefficient (e.g., Bradshaw and Anderson, 2001) or, more commonly, squared-chord distances (e.g., Kauppila and Valpola, 2003; Tibby *et al.*, 2003; Bennion *et al.*, 2004). In a comparison of the two methods by Ryves *et al.* (2002) found very little difference. Authors differ in terms of what threshold of DC indicates good or poor analogues. Both Bradshaw and Anderson (2001) and Ryves *et al.* (2002) assign a simple 50/50 valuation where a poor analogue is greater than 50% of the DC and a good analogue is less than 50%. Most other authors assign a far stricter criterion with the most commonly assigned threshold being that DCs greater than the lowest 10<sup>th</sup> percentile equate to a poor or very poor analogue (Laird *et al.*, 1998; Kauppila and Valpola, 2003; Tibby *et al.*, 2003), though some go further suggesting that only values less than the lowest 5<sup>th</sup> percentile (Bennion *et al.*, 2004), or even the lowest 2<sup>nd</sup> percentile (Rioual *et al.*, 2007), is representative of a good analogue.

In this study, fossil samples with a minimum squared-chord distance less than the 5<sup>th</sup> percentile of the DCs of the calibration set were considered to have very good modern analogues and those that greater than the 5<sup>th</sup> and less than the 10<sup>th</sup> percentile were considered to have good modern analogues. Those that fell below this level were considered to have poor modern analogues. The percentiles of the training set dissimilarities are presented in Table 5.26

Table 5.26: Percentiles of the training set dissimilarities

Percentiles	Squared-chord distance
1 <sup>st</sup> percentile	49.38
2 <sup>nd</sup> percentile	67.88
5 <sup>th</sup> percentile	99.97
10 <sup>th</sup> percentile	131.49
20 <sup>th</sup> percentile	163.58

Figure 5.43 illustrates analogue matching results for Lake Elingamite core LE1 and how it corresponds to the diatom-inferred conductivity reconstruction. As can be seen, the entire reconstruction has either good, or very good, modern analogues. Indeed, by far the majority of the core has very good modern analogues, indicating that palaeoconductivity interpretations made from this core can be undertaken with some confidence.

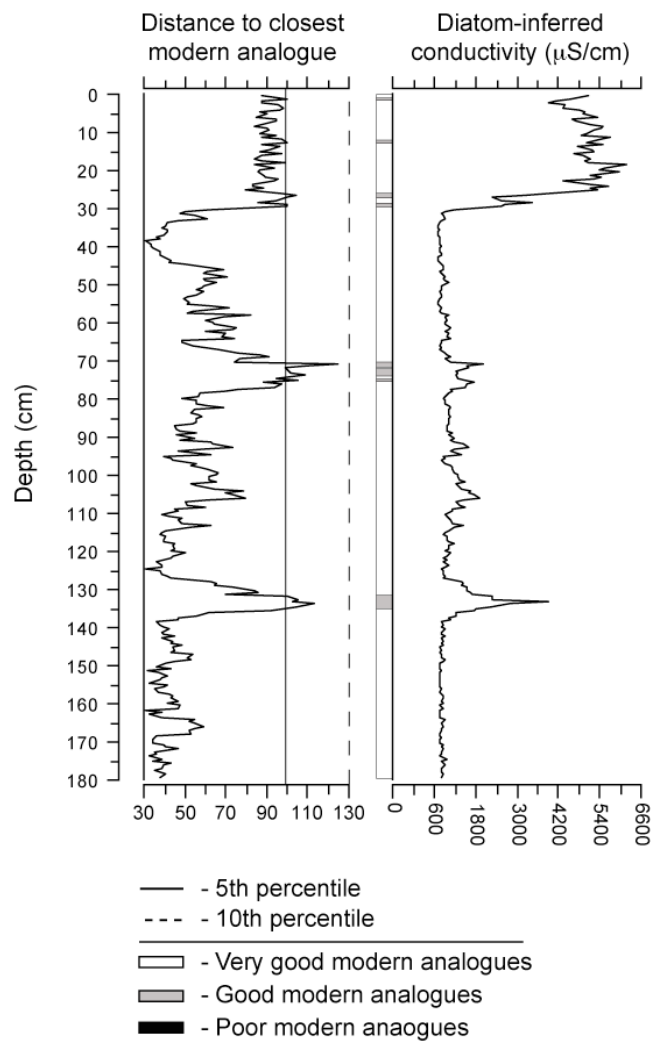


Figure 5.43: Modern analogues and diatom-inferred conductivity in core LE1.

The results of applying the MAT to core LSFS from Lake Surprise are presented in Figure 5.44. Again, the majority of the core has good or very good modern analogues though a small portion is considered to have poor analogues in the modern training set.

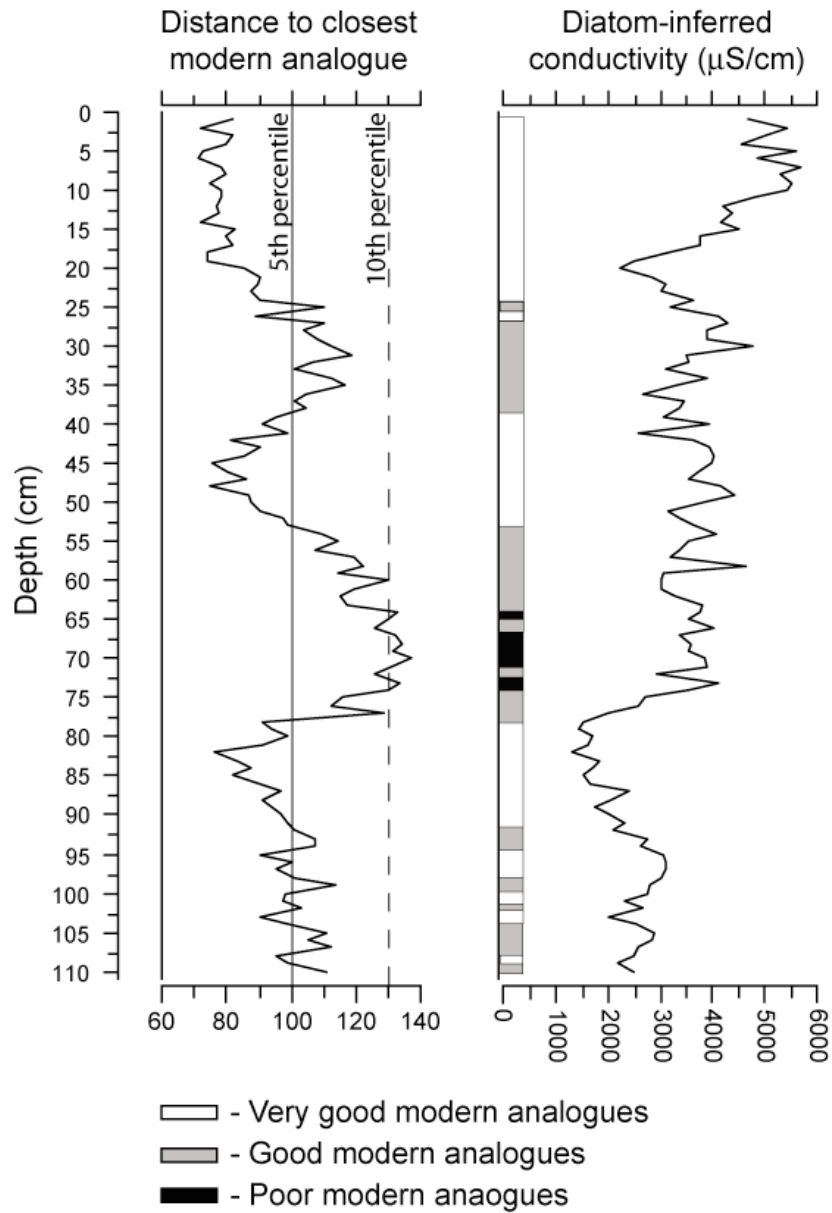


Figure 5.44: Modern analogues and diatom-inferred conductivity in core LSFS

The modern analogues of core LST1 from Lake Surprise are presented in Figure 5.45. As can be seen, much of the core has poor analogues, though over half has good or very good analogues. 21% of samples have very good analogues, 35% have good analogues,

while 44% have poor analogues. Interpretation of the palaeoconductivity record from some portions of the Lake Surprise core should therefore be undertaken with caution.

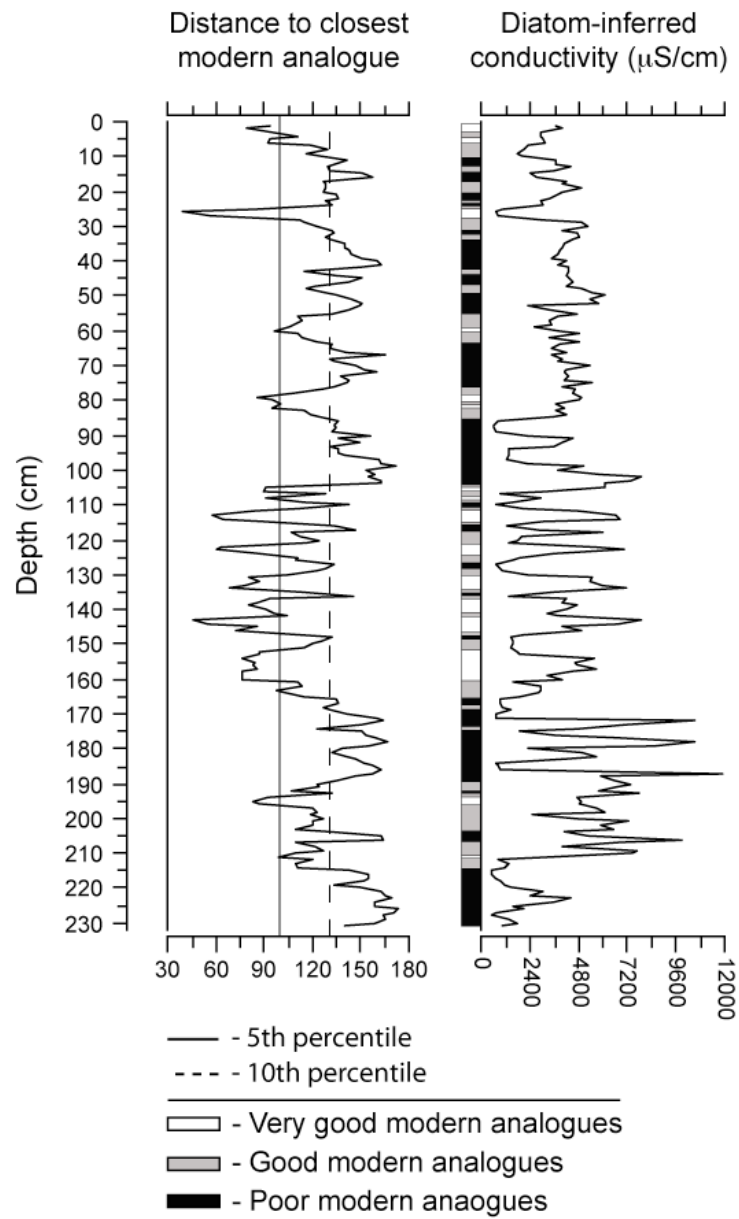


Figure 5.45: Modern analogues and diatom-inferred conductivity in core LST1

## CHAPTER 6 – DISCUSSION

In this chapter, the results of the analysis of the cores from Lakes Elingamite and Surprise will be discussed in light of the evidence presented above and the literature review presented in Chapter 2. The interpretation from both study sites is based on a chronology which is established in Section 6.1. In light of the results of correlating the Lake Surprise cores presented in Section 5.3, Section 6.1.2.1 finalises the merging of the two Lake Surprise cores and establishes a single record from the lake. In order to assess the responsiveness of the lakes to climate forcing, recent reconstructions from the study sites are compared against historical data in Section 6.2. Once it has been demonstrated that both lakes are climatically sensitive, the records from both lakes are compared to identify patterns of divergence and coherence between sites (Section 6.3). Prior to discussing the climate history of the study region, an interpretive framework is developed in order to address both site-specific and general issues which may hamper analysis (Section 6.4). This framework provides the foundation for interpretation of results and the reconstruction of palaeoclimates in western Victoria, which will constitute the remainder of the chapter.

### 6.1 Establishing lake chronologies

In order to attain the stated aims of the project and provide insight into the climate history of western Victoria, it is necessary to place the results presented in Chapter 5 in a firm chronological context. However, as outlined in Chapter 4, several problems were identified with the dating from both lakes. The following discussion will establish a final chronology for both study sites by drawing together all available evidence to develop the most reasonable chronological model. Each site will be discussed individually. In each case, the chronology of the most recent sediments will be discussed first in order to provide the means to quantify any age offset that may affect the radiocarbon dates due to the presence of old carbon.

#### *6.1.1 Lake Elingamite chronology*

As outlined in Chapters 4 and 5, determining a chronology for the Lake Elingamite record was not straightforward. Recent sediments were dated using gamma spectrometry in the first instance, followed by the dating of further samples by alpha spectrometry in an attempt to extend the recent chronology. All  $^{210}\text{Pb}$  results were

analysed using the modified CIC method of Brugam (1978). The CRS method (Appleby and Oldfield, 1978) could not be used to interpret  $^{210}\text{Pb}$  results due to an absence of sediment bulk density data because of desiccation of core samples after subsampling into 0.5 cm slices (see Section 4.2.1.2.1).

The profile of unsupported  $^{210}\text{Pb}$  activity obtained by gamma spectrometry (Figure 5.3) is markedly different to that obtained by alpha spectrometry (Figure 5.6). The gamma spectrometry profile has an arched appearance and, as a result, interpretation of the data consisted of extrapolating the calculated sedimentation rate of the lower half of the samples to the core surface. The  $^{210}\text{Pb}$  versus depth relationship derived via alpha spectrometry has a more classical, linear appearance, however the derived ages are significantly younger than those derived through gamma spectrometry. The differences between the plotted  $^{210}\text{Pb}$  data, and the derived ages, highlights the need for independent evaluation of the derived age-depth relationships. Unfortunately, the activities of  $^{137}\text{Cs}$  in this core do not show any indication of a period of atomic testing (Table 5.1) which raised concerns that  $^{137}\text{Cs}$  may be mobile in the sediment (Atun Zawadzki, pers. comm.) and thereby excluded this as a method of validating the  $^{210}\text{Pb}$  derived chronologies. Other studies have invoked the arrival of exotic *Pinus* pollen in the sedimentary record as a means to determine chronologies (e.g., Dodson *et al.*, 1994) and this is a possibility for Lake Elingamite. A major shift in the diatom record occurs concurrently with the arrival of *Pinus* in the sediments at a depth of 56 cm, suggesting a period of significant landscape alteration as would be expected by the arrival of European settlers. Yet, the alpha spectrometry-derived age estimates of this change conflict with the historical record.

The key alpha spectrometry dates that challenge the creation of a clear age-depth relationship are from samples IO86 (50-52 cm) and IO87 (60-62 cm). The ages for these samples were calculated to be  $45 \pm 4$  years and  $66 \pm 7$  years, respectively. Exotic *Pinus* pollen first appears in the record roughly midway between these two samples at 56 cm and, according to these dates, this is approximately  $55 \pm 5$  years before 2005 (the reference date – see Table 5.1), or 1950 AD. However, European settlement began in the region over a century earlier, around 1840 AD, and pine windbreaks were soon established by the early settlers (Kiddle, 1967; Powell, 1970). The arrival of *Pinus* pollen in western Victorian sedimentary records has generally been inferred to represent a date of *ca.*1850 AD (Dodson and Mooney, 2002; Tibby *et al.*, 2006), as is the case in

a study from Cobrico Crater approximately 5 km to the north of Lake Elingamite (Dodson *et al.*, 1994). However, care must be taken in applying such an all-encompassing appraisal as several studies have shown the first appearance of *Pinus* to post-date European induced vegetation changes by as much as 50 years (e.g., Mooney *et al.*, 2000; Gell *et al.*, 1993). Even so, a lag of 50 years in this instance would suggest an age around 1900 AD. This date still falls well short of that derived by alpha spectrometry (1950 AD) and supports the notion that these calculated dates are younger than the real age of the sediment.

It therefore appears that, despite the abnormal  $^{210}\text{Pb}$  profile, the gamma spectrometry  $^{210}\text{Pb}$  data provide the more accurate dates for the upper sediments. Sample H739 is the lowest gamma spectrometry date, at a depth of 45 cm, and has a calculated age of 101 years. Continuing the inferred sedimentation rate of 0.45 cm/yr to a depth of 56 cm returns an extrapolated age of 125 years, or 1880 AD, which is sufficiently consistent with the probable arrival of *Pinus* pollen to constitute a more plausible scenario than that derived from the alpha spectrometry dates.

However, while it can be reasonably argued that 1880 AD represents a realistic estimate of the arrival of *Pinus*, evident at a depth of 56 cm, this is at odds with the radiocarbon dates which were taken from the same sedimentary level, which are notably older. This raises the prospect that a reservoir effect is present in Lake Elingamite. This is not surprising, given that this lake is maintained by groundwater inflow and is underlain with limestone (Mann, 1991), both of which can act as sources of old carbon (Cohen, 2003).

The problem of radiocarbon samples recording unexpectedly young or old ages is not uncommon (e.g., Pazdur *et al.*, 1995; Wohlfarth *et al.*, 1998; Turney *et al.*, 2000). Results from other volcanic lakes in western Victoria suggest that radiocarbon dates may often return ages that are between 200 (Barton and Polach, 1980) and 450 years (Barton and Barbetti, 1982) too old. Indeed, closer inspection of some regional studies confirms that a reservoir effect is not uncommon. Dodson (1974) for instance, returns a date of  $765 \pm 135$  radiocarbon years BP approximately 5 cm below the appearance of exotic *Pinus* pollen in the Lake Keilambete record. Barton and Barbetti (1982:375) suggested that dates from Lake Keilambete are “about 450 years too old.” From the same lake, Mooney (1997) re-examined and calibrated all previously reported

radiocarbon dates and calculated a surface sediment age of 265 cal. BP. Similarly, from Lake Bullenmerri, Barton and Polach (1980:736) calculate that there is a “25 percent probability of a particular  $^{14}\text{C}$  age being in error by more than 20 percent.” Results from Cobrico Crater (Dodson *et al.*, 1994) have an even greater apparent error, indicating a radiocarbon age of  $1710 \pm 60$  BP from 15 – 20 cm depth, while exotic *Pinus* pollen first appears at 18 cm. The decision to date pollen concentrates, and not sedimentary organic matter, was taken in an effort to avoid this problem, based on the assumption that the majority of the pollen in the sediment was of terrestrial origin.

Samples Wk-17821 and Wk-17822 were taken at the same level as the first appearance of *Pinus* in an effort to, firstly, quantify dating errors from the reservoir effect and, secondly, to gauge the impact of laboratory methods which destroyed the pollen during the sample preparation process (Section 4.2.5.1). The samples were sent to a different laboratory and underwent identical pre-treatment before sample Wk-17821 was subjected to further treatment in order to isolate and concentrate the pollen grains through heavy liquid separation. Abundant pollen was present in the sample (Fiona Petchey pers. comm.) and it was this pollen concentrate that was subjected to dating. It was assumed, given the small area of littoral vegetation compared to the total lake size, that the pollen concentrate would consist primarily of terrestrial pollen, and would thereby present an age younger than that of the paired sample consisting primarily of autochthonous material. Even though the laboratory was unable to report the type of pollen present, the fact that the concentrated pollen sample returned an age 557 radiocarbon years (443 Cal.yr) older than the paired sample was still surprising. The reason for this is not entirely clear, though Kilian *et al.* (2002) provide a possible explanation. In their study of a varved lake in Poland, the authors also found that ages derived from concentrated pollen samples were older than expected. They explained this by stating that;

*“Only the walls of pollen grains are preserved. These walls have openings (preformed apertures and ruptures), and therefore pollen have a very high surface/volume ratio compared to macroscopic remains like seeds. Pollen walls have a porous structure that can accommodate either adsorbed organic molecules or act as a nucleation space for carbonate deposition. Due to the poor wettability of the pollen pore space, standard pretreatment may not be sufficient to remove such allochthonous components (Kilian et al., 2002:25).”*



Therefore, the act of concentrating pollen may have also inadvertently concentrated the carbonate responsible for returning old ages. In this sense, the organic content of the other dated samples may have been modulating the effect of the pollen, potentially explaining the markedly younger age of sample Wk-17822, which still contained the organic matrix remaining after pre-treatment for pollen. Some support for this hypothesis is evident from the dates taken at 146 cm (OZ117), which consisted of the organic matrix remaining after pollen preparation, and 147-149 cm (Wk-17870), which was dried organic sediment. The calibrated dates, derived by taking the midpoint between the two-sigma age range with the largest relative area of probability, are very similar (71 years difference) and, even though they are stratigraphically 'reversed', their two-sigma age ranges overlap by 51 years (Table 5.6). The fact that two samples with 'diluted' pollen plot so closely, while two samples consisting of concentrated pollen and an organics/pollen matrix plot so distant to each other, lends weight to the argument that the presence of pollen may be the cause of the extraneous dates.

In terms of quantifying the impact of old carbon on the dates, the older concentrated pollen sample (Wk-17821) was therefore ignored, while the age of the younger 'organic matrix' sample (Wk-17822), 1333 AD, was subtracted from the extrapolated gamma spectrometry  $^{210}\text{Pb}$  age, 1880 AD. This results in an age difference of 547 years. While other studies have employed this same method of age correction (e.g., Stanley and De Deckker, 2002), there are limitations associated with the application of such a blanket application. In particular, the rate, source and age of the old carbon may have changed over time (Geyh *et al.*, 1998). However, as these factors are unable to be calculated, this is the only means of correcting the dates (in the absence of further dating options).

The final (preferred) age-depth model for Lake Elingamite is presented in Figure 6.1 and incorporates the 547 year correction. It is, however, recognised that a number of caveats should be considered in the application of this final chronology. While applying a point-to-point interpolation approach between the calibrated dates is commonly undertaken in palaeolimnological studies, especially where chronological data are limited (Cohen, 2003), there are inherent weaknesses in this technique. Firstly, it assumes that sedimentation rates change at exactly the same point as the dated sample. Secondly, it assumes that sedimentation rates are constant between the dated samples. Thirdly, it forces the age-depth model to pass through set 'points', even though these are only a probable location of the 'true' dates (Telford *et al.*, 2004) and, finally, there is

a strong likelihood that errors exist in the interpolated ages when there are few data points (Cohen, 2003; Telford *et al.*, 2004). The interpretation presented here is undertaken in full appreciation of the limitations of the process, yet the need to place the derived results from the lake in an historical context encourages a pragmatic approach to be taken. Therefore, while the likelihood of errors in the presented interpretation is acknowledged, the approach that has been taken is the most practicable.

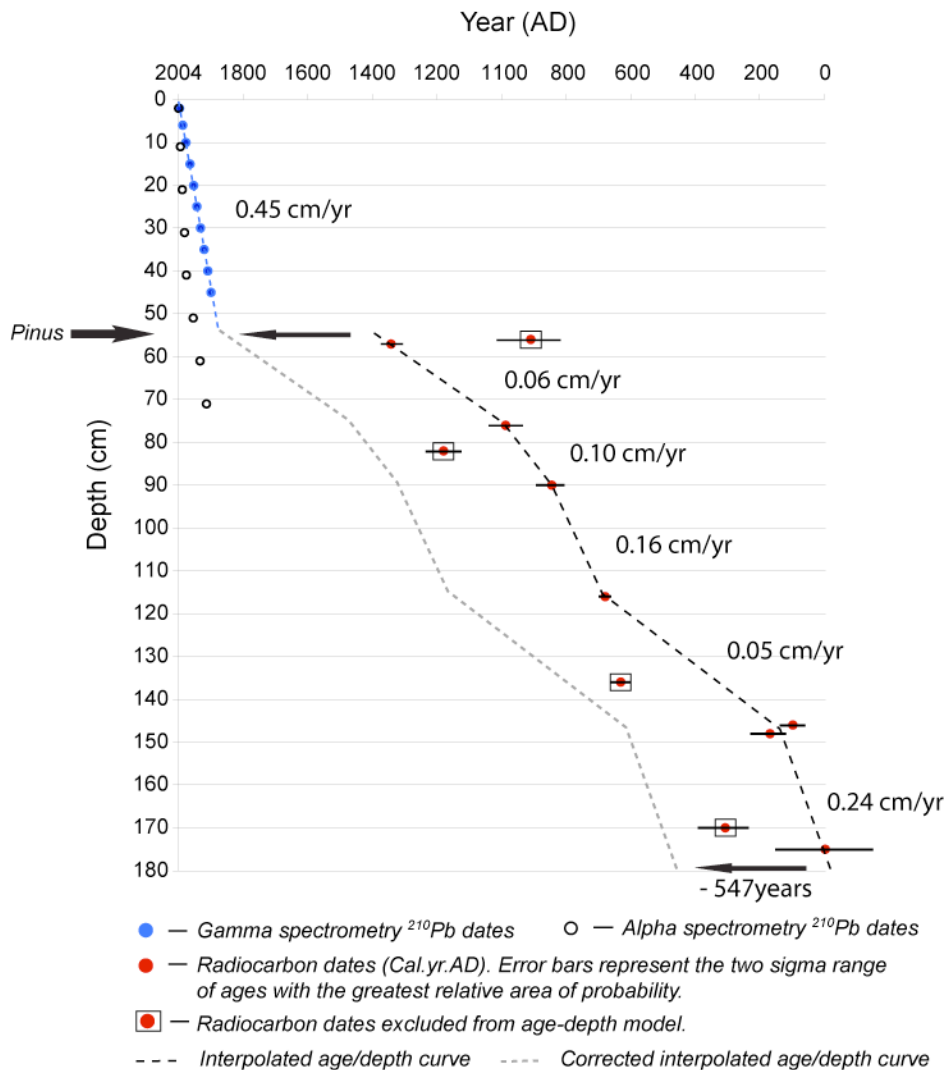


Figure 6.1: The age-depth model for Lake Elingamite showing corrected ages and inferred sedimentation rates

The linear interpolation presented here is a parsimonious approach to developing the age-depth profile. Four dates were ignored in the development of the profile and are identified in Figure 6.1 above. The first date at 56 cm was that from the concentrated

pollen sample and the reason for its exclusion has been discussed above. The next two dates down core were excluded as the inferred sedimentation rate between them, and the next logical date above, would have been very similar to the post-European sedimentation rate. Australian palaeolimnological studies consistently find an increased sedimentation rate following the arrival of Europeans and the landscape alterations that followed (Dodson and Mooney, 2002; Gell *et al.*, 2005), with thirty to forty-fold increases in sedimentation rates commonly found for riverine wetlands (e.g., Leahy *et al.*, 2005; MacGregor *et al.*, 2005). A more moderate increase could be expected from crater lakes due to their smaller catchments, though data is sparse. Mooney and Dodson (2001) provide the only conclusive evaluation of post-European sedimentation rates from crater lake systems in the study area, finding a 2½ fold increase in Lake Keilambete. Therefore, a pre-European sedimentation rate in Lake Elingamite that is comparable to a post-European rate would represent a period of massive landscape alteration and as such, provides support for the exclusion of these two dates.

The final excluded date, at a depth of 170 cm, was omitted as it appears to be a clear outlier in the model, especially given the greater confidence in the two dates immediately above in the core, which plot very near to each other. A counter argument would be that these two dates may represent the same period of redistribution of old sediment, however, this does not appear to be the case. The sediment of core LE1 was homogenous in nature and colour. Furthermore, the bathymetry of Lake Elingamite indicates a very broad, flat-bottomed lake (Figure 3.13), making large-scale sediment slumping unlikely. Therefore, more confidence is placed in these dates than the one immediately beneath them stratigraphically. Extrapolating the finalised age-depth profile to the base of the core indicates that the Lake Elingamite record spans a period of 1483 years with each sample in the post-European phase equating to approximately 2.3 years and each sample in the pre-European phase representing an average of 5.5 years. Having settled on a preferred chronology, a diatom stratigraphy can be developed which illustrates the fluctuations in the relative abundance of the main species in a temporal context (Figure 6.2).

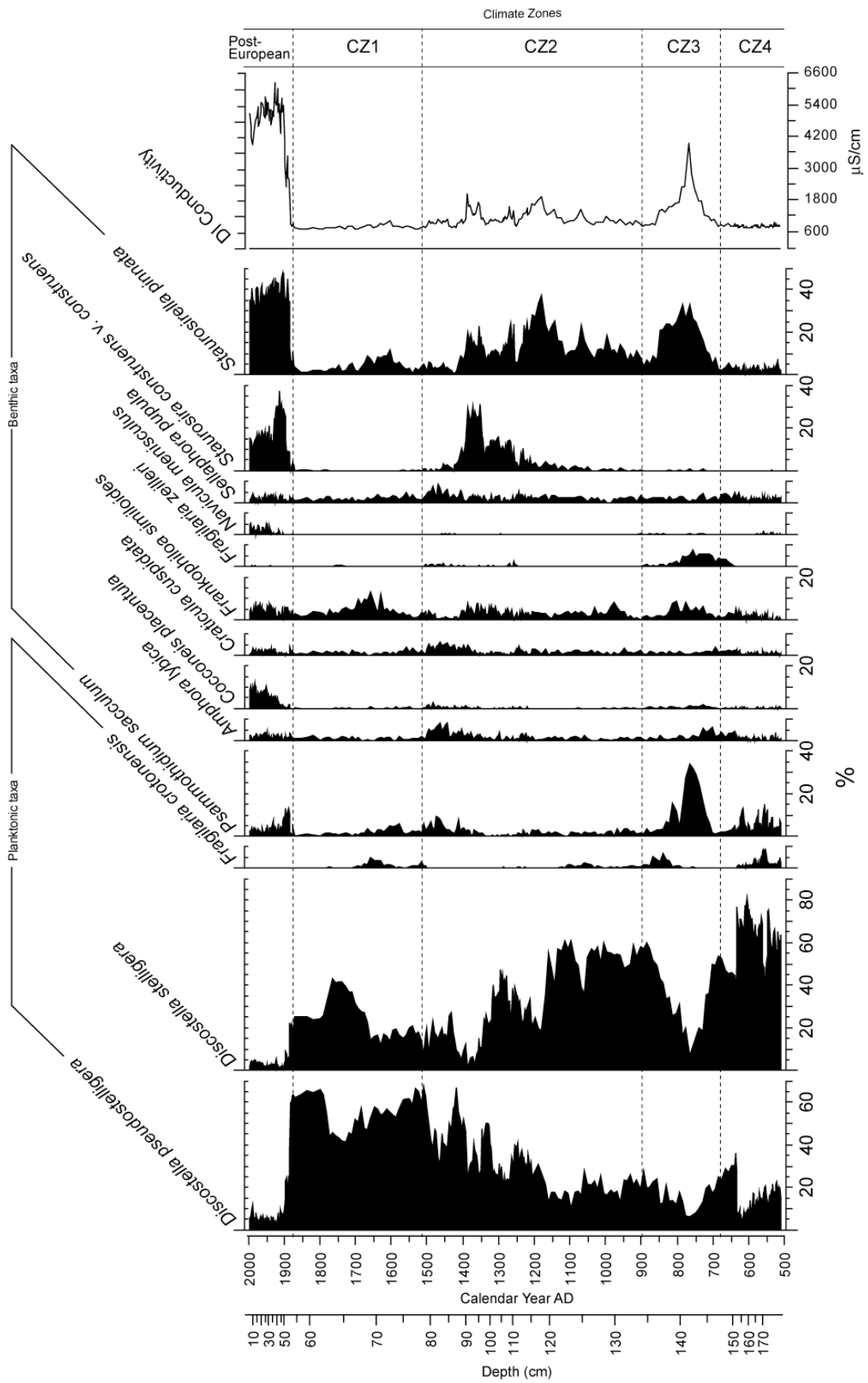


Figure 6.2: Lake Elingamite diatoms and diatom-inferred (DI) conductivity, plotted by age. Only species with a relative abundance > 5% in at least one sample are included. Climate zones are established in Section 6.5 and included for reference.

### 6.1.2 Lake Surprise chronology

The Lake Surprise sedimentary sequences were dated using  $^{210}\text{Pb}$  for the more recent portion of the record and  $^{14}\text{C}$  AMS for the older sediments. Independent chronological evidence is also provided by pollen analysis (Figure 3.18; Chris White, unpublished data) and sedimentary mercury data (Figure 3.20; Jacqui Hellings, unpublished data).

The  $^{210}\text{Pb}$  data was analysed using both the modified Constant Initial Concentration (CIC) model (Brugam, 1978) and the Constant Rate of Supply (CRS) method (Appleby and Oldfield, 1978). Using the CIC model, the excess  $^{210}\text{Pb}$  activity versus depth plot (Figure 5.14), indicates that samples from depths of 91 cm, 81 cm and 71 cm had very low levels of excess  $^{210}\text{Pb}$ . This was interpreted as being indicative of background levels of excess  $^{210}\text{Pb}$  and was not used to derive age estimates (Atun Zawadzki, pers. comm.). The samples were re-analysed using the Constant Rate of Supply (CRS) method (Appleby and Oldfield, 1978), resulting in age determinations for the samples at 71 cm and 81 cm (Table 5.9).

An examination of the CIC-derived ages and sedimentation rates (Figure 5.16) indicates a period of rapid sediment deposition from 51 cm to 31 cm, while sedimentation rates from 61-51 cm, and 31-0 cm, were much more gradual, being 0.27 and 0.32 cm/yr respectively (Table 5.8). Similarly, the two calibrated  $^{14}\text{C}$  dates, from 81 cm and 104 cm of core LSFS, indicate a comparable sedimentation rate of 0.28 cm/yr. This suggests that, apart from a period of very rapid sedimentation rate from 51 cm to 31 cm, sediments in this record have been deposited at a reasonably constant rate. However, calculating the sedimentation rate between the uppermost  $^{14}\text{C}$  date at 81 cm (532 cal. BP) and the bottom  $^{210}\text{Pb}$  date at 61 cm (~130 yrs BP), reveals a period where the sedimentation rate was only 0.05 cm/yr. There is no evidence in the stratigraphy, diatom or pollen records to support a period of excessively slow sedimentation, nor is any evidence present in analysis of the sediment where results for bulk density, water content, organic content and carbonate content over this period show no obvious change. Given that the calculated sedimentation rates, both above and below this period, are practically identical, the prospect is raised that the radiocarbon dates are 'too old'.

The calculated CRS ages (Table 5.9) also challenge the  $^{14}\text{C}$  age determinations as, in this instance, an age could be calculated from the same sedimentary level as the uppermost  $^{14}\text{C}$  date at 81 cm: approximately 119 years before present (Figure 5.17).

This is markedly different to the  $^{14}\text{C}$ -derived 532 cal. BP age, and raises the notion that, as was the case in Lake Elingamite, radiocarbon dates from Lake Surprise are being affected by the presence of old carbon.

The most likely source of old carbon is the local groundwater which flows through the underlying Tertiary marls and limestones (Douglas, 1982) and maintains the lake. However, the degree to which this alters the radiocarbon dates needed to be quantified by an independent marker, in order for the dates to be corrected (*sensu* Shotton, 1972). The CIC-derived chronology does not provide a date at a depth of 81 cm to correspond with the upper  $^{14}\text{C}$  sample, though an extrapolation of the age-depth profile suggests an approximate 307 year offset (Figure 6.3). In contrast, the CRS-derived chronology did provide an age determination for the sample at 81 cm;  $118 \pm 1.3$  years, suggesting an approximate 413 year offset. Given the discrepancy between the CIC- and CRS-derived ages, there is an obvious requirement to determine which method provides the more accurate chronology. As samples from core LSFS was analysed using alpha-spectrometry though, no  $^{137}\text{Cs}$  data exists which may act as an independent marker to validate the CIC or CRS chronologies. Therefore, results were compared against other chronostratigraphic markers.

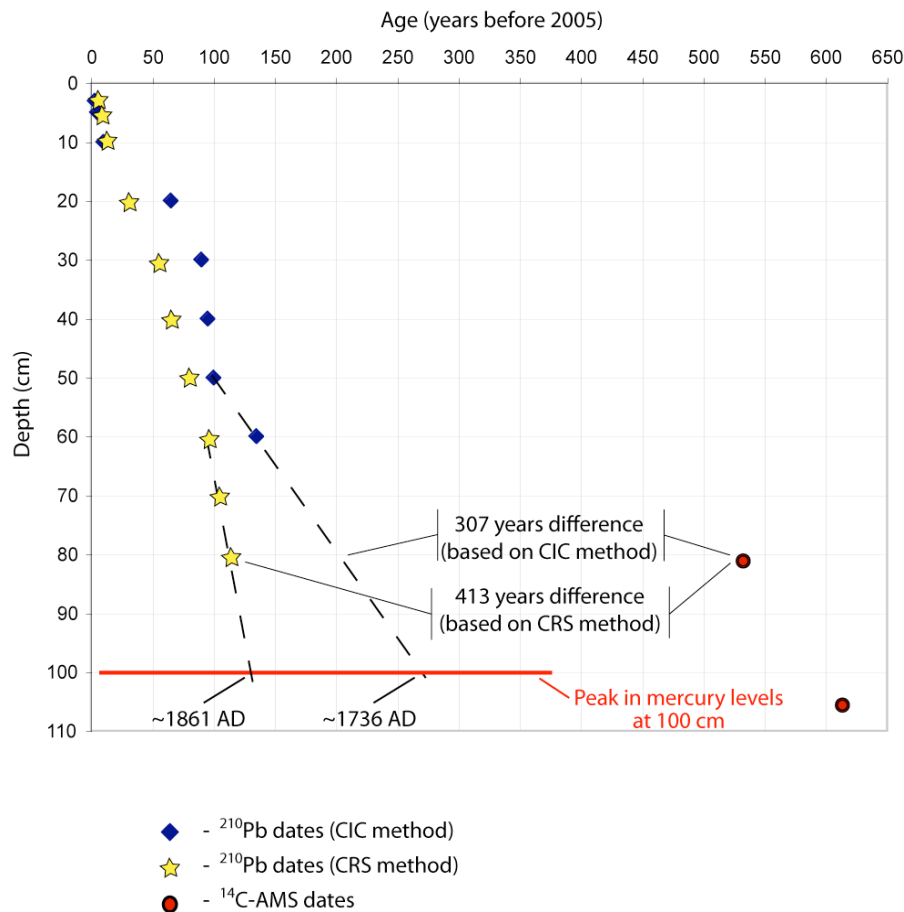


Figure 6.3: Determinants in quantifying the effect of old carbon in the Lake Surprise catchment. All dates from Core LSF5.

The appearance of exotic *Pinus* pollen is commonly used in Australian palaeo research as such a marker and, in western Victoria, is generally interpreted as indicating a period somewhere between 1850 AD (Tibby *et al.*, 2006) and 1870 AD (Leahy *et al.*, 2005). While pollen analysis of core LSF5 indicates the first appearance of *Pinus* occurred at a depth of 90 cm (Figure 3.18), only single grains were found in sporadic samples below 60 cm (Chris White, unpublished data). While no *Pinus* is apparent below 90 cm, the possibility that the low levels of *Pinus* from 60 cm to 90 cm are due to contamination cannot be discounted. Even so, extrapolation of CIC-derived age estimates to a depth of 90 cm indicates an approximate age of 1765 AD, which pre-dates European arrival on the continent, compared to an extrapolated age of 1875 AD derived from the CRS data, which is far more plausible. Despite the obvious argument in favour of the CRS-derived data though, the use of exotic *Pinus* as an age determinant for this core was disregarded, primarily because any age determination would ultimately be based upon the presence of a single pollen grain, which could quite reasonably be interpreted as contamination.

An alternative chronological marker exists in the levels of mercury contained in the sediments of Lake Surprise, which abruptly increase at a depth of 100 cm (Figure 3.20). The precise source of this mercury is unknown (Jacqui Hellings, pers. comm.), however, as mercury is easily transported in the atmosphere (Jackson, 1997; Schroeder and Munthe, 1998), it is reasonable to suppose that it emanates from the goldfields of western Victoria (see Figure 3.1), where the use of mercury in the gold mining process was widespread during the gold rush of the 1850s and 1860s (Bycroft *et al.*, 1982; Churchill *et al.*, 2004), or from the burning of fossil fuels such as coal (Slemr and Langer, 1992). Extrapolating the CIC-derived age estimates to a depth of 100 cm indicates an approximate age of 1736 AD, which, again, is incongruous with the historical record. In contrast, extrapolation of CRS-derived age estimates indicates an approximate age of 1861 AD at 100 cm, which correlates well with historical events. A comparison of all lines of evidence therefore suggests that the CRS dating model provides more logical results than the CIC model for core LSFS.

The calculated CRS age at 81 cm was  $118.9 \pm 1.3$  years. Contrasting this age against the calibrated  $^{14}\text{C}$  date of 532 years from the same sedimentary level suggests the age offset from old carbon is 413 years. Using this information, all calibrated  $^{14}\text{C}$  dates from cores LSFS and LST1 were corrected by subtracting 413 years. As with the Lake Elingamite record, the calculated age difference was deducted from all other calibrated  $^{14}\text{C}$  dates in the record and was undertaken with recognition of the same limitations and caveats outlined above.

Five  $^{14}\text{C}$  ages were ascertained from core LST1 and the corrected, calibrated dates are illustrated in Figure 6.4. While there is a strong age-depth relationship present, with ages generally increasing with depth, two dates plot as 'reversals' where they are older than the date directly below them. As Figure 6.4 illustrates, there are a number of alternative interpretations of the age-depth relationship and, therefore, each needs to be evaluated prior to a final chronology being determined. With the exception of option 3, each option includes three dates and, therefore, two calculated sedimentation rates. These are provided in Table 6.1.



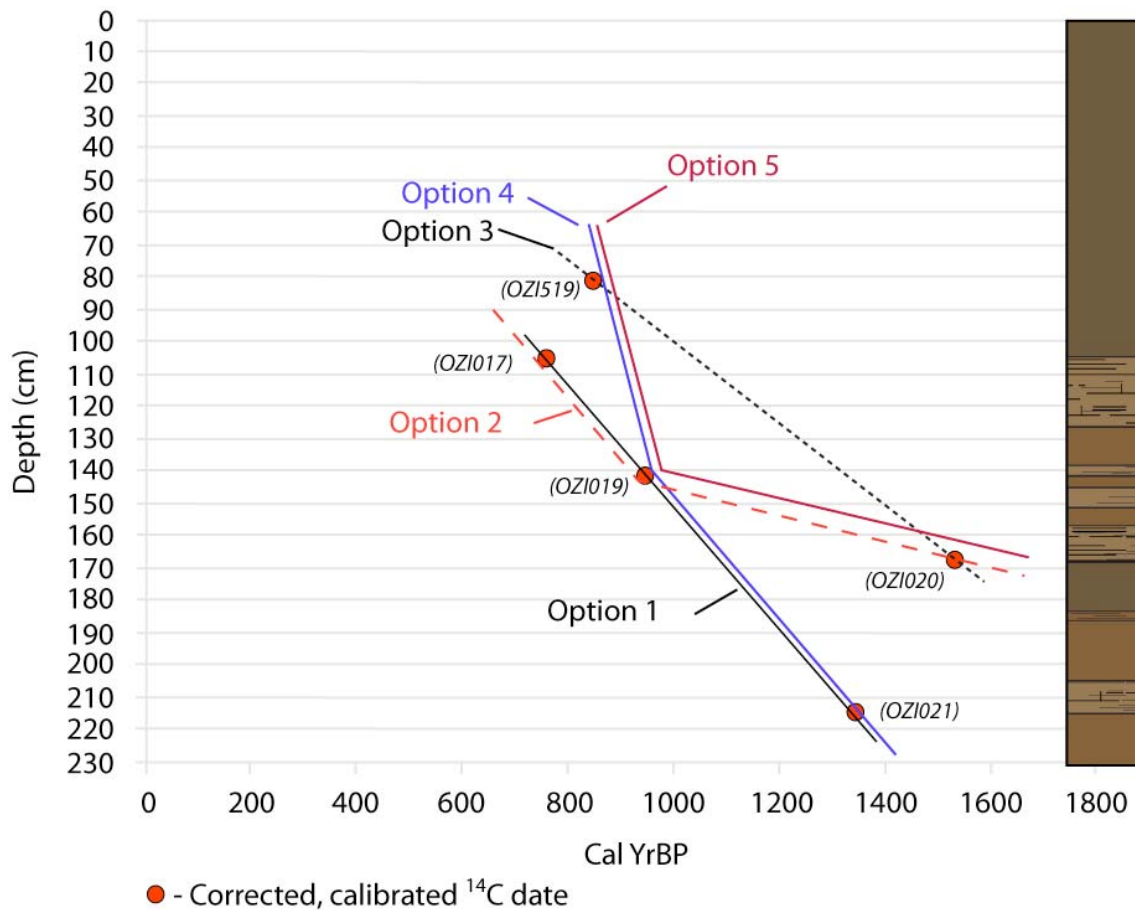


Figure 6.4: Potential sediment age-depth models for interpreting core LST1 chronology with laboratory codes and core lithology (see Figure 5.19 for key) provided for reference. Note: dates have been corrected by subtracting 413 years.

Table 6.1: Calculated sedimentation rates for core LST1 based on options presented in Figure 6.4

Option	Sample depths	Sedimentation Rate (cm/yr)	Sample depths	Sedimentation Rate (cm/yr)
Option 1	104 - 142	0.21	142 - 215	0.19
Option 2	104 - 142	0.21	142 - 168	0.05
Option 3	81 - 168	0.13		
Option 4	81 - 142	0.62	142 - 215	0.19
Option 5	81 - 142	0.62	142 - 168	0.05

Options 4 and 5 were dismissed due to the very high calculated sedimentation rate between 142 cm and 81 cm of 0.62 cm/yr, which is comparable to the average post-European sedimentation rate of 0.68 cm/yr (calculated from the CRS model). As discussed previously in regard to the Lake Elingamite record, for this to occur would suggest a level of ecosystem disturbance similar to that which occurred after the arrival of Europeans. As this is a highly unlikely scenario, options 4 and 5 can be confidently rejected.

Option 2 indicates a sedimentation rate of 0.05 cm/yr from core depths of 168 cm to 142 cm, followed by an increase in sedimentation rate to 0.21 cm/yr from 142 cm to 104 cm. This represents a four-fold increase in the rate of sediment deposition. From 168 – 142 cm, and 142 – 104 cm, there are several periods in which carbonate-rich sediments have been laid down, interspersed by periods of organic-rich lake mud deposition. These periods are also reflected in the sediment analysis with large scale variability evident in all the measured variables. This suggests a highly variable environment during both these sections of the core and it could be assumed that sedimentation rates would have fluctuated accordingly. It is also possible though, that these oscillations were occurring across a solubility threshold and resulted in little change to the sedimentation rate. In either case, the calculated sedimentation rates presented here only represent an average over the period between two distinct dates and are therefore a ‘smoothing’ of the actual sedimentation rates. As such, it is difficult to conceive that these averages would be so divergent that one would be four times greater than the other. For this reason, option 2 was also disregarded.

Therefore, options 1 and 3 seem equally probable scenarios on which to base the interpretation of the record. The sedimentation rate for option 1 was 0.21 cm/yr and 0.19 cm/yr, while for option 3 it was 0.13 cm/yr. Either of these could reasonably be argued to be the best scenario as both inferred sedimentation rates could be expected to be within the range of possibility. Of these two options though, option 1 appears to be the most robust as it includes three data points compared to only two data points in option 3. The sedimentation rates for option 1 are also similar to the sedimentation rate between the two calibrated  $^{14}\text{C}$  dates at the bottom of Core LSFS (0.29 cm/yr). The fact that they are slightly lower could be due to sediment compaction, or because the corrected LSFS dates fall within the early European period where slightly higher sedimentation rates could be expected. Option 1 therefore, is the scenario selected in all

further analyses of the Lake Surprise record. Extrapolating the corrected calibrated  $^{14}\text{C}$  to the base of the core provides a basal age of 579 AD with each sample in the post-European phase equating to approximately 1.5 years and each sample in the pre-European phase representing an average of 4.9 years.

As with the Lake Elingamite record, the age-depth model was developed using linear interpolation between dates. Despite the limitations and potential for error using linear interpolation, it has been found to be at least as accurate, in terms of predicted errors, than other methods of interpolation when only a few data points are available (Telford *et al.*, 2004) and can be considered to be a good general solution that is “rarely unacceptably wrong” (Bennett and Fuller, 2002:427).

#### *6.1.2.1 The full chronology from the merged Lake Surprise cores*

In Section 5.3, the process of correlating cores LSFS and LST1 biostratigraphically using the diatom record, and also via their chronologies, was discussed. Using these methods, it was demonstrated that the most likely scenario was that the two cores did not overlap, yet extrapolation of the chronology-inferred sedimentation rates indicated that the gap between the cores was very small (approximately 3 cm, *ca.* 15 years). With this gap being identified, and as a final chronology for both cores has now been established, as discussed above, the full Lake Surprise chronostratigraphy can be presented (Figure 6.5). Establishing the final chronology also enables the diatom record to be plotted by age rather than depth, thus providing a more realistic representation of the variability of results such as DI conductivity and sediment analyses. Figure 6.6 illustrates the relative abundance of the major diatom species. As opposed to Figures 5.31 and 5.32 presented earlier, which were plotted by depth and indicated a dramatic shift from a highly variable record in LST1 to a far more stable record in LSFS, plotting by age demonstrates that, while this shift is still obvious, it is not so dramatic. The full Lake Surprise diatom-inferred conductivity record is illustrated in Figure 6.7.

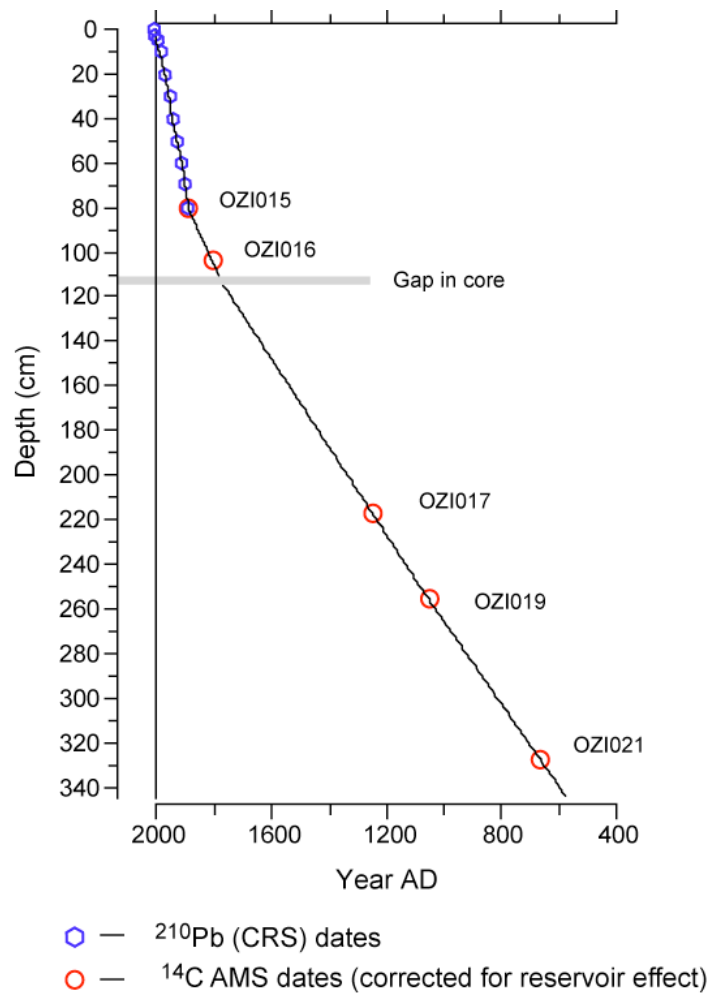


Figure 6.5: Chronostratigraphy of the full Lake Surprise record

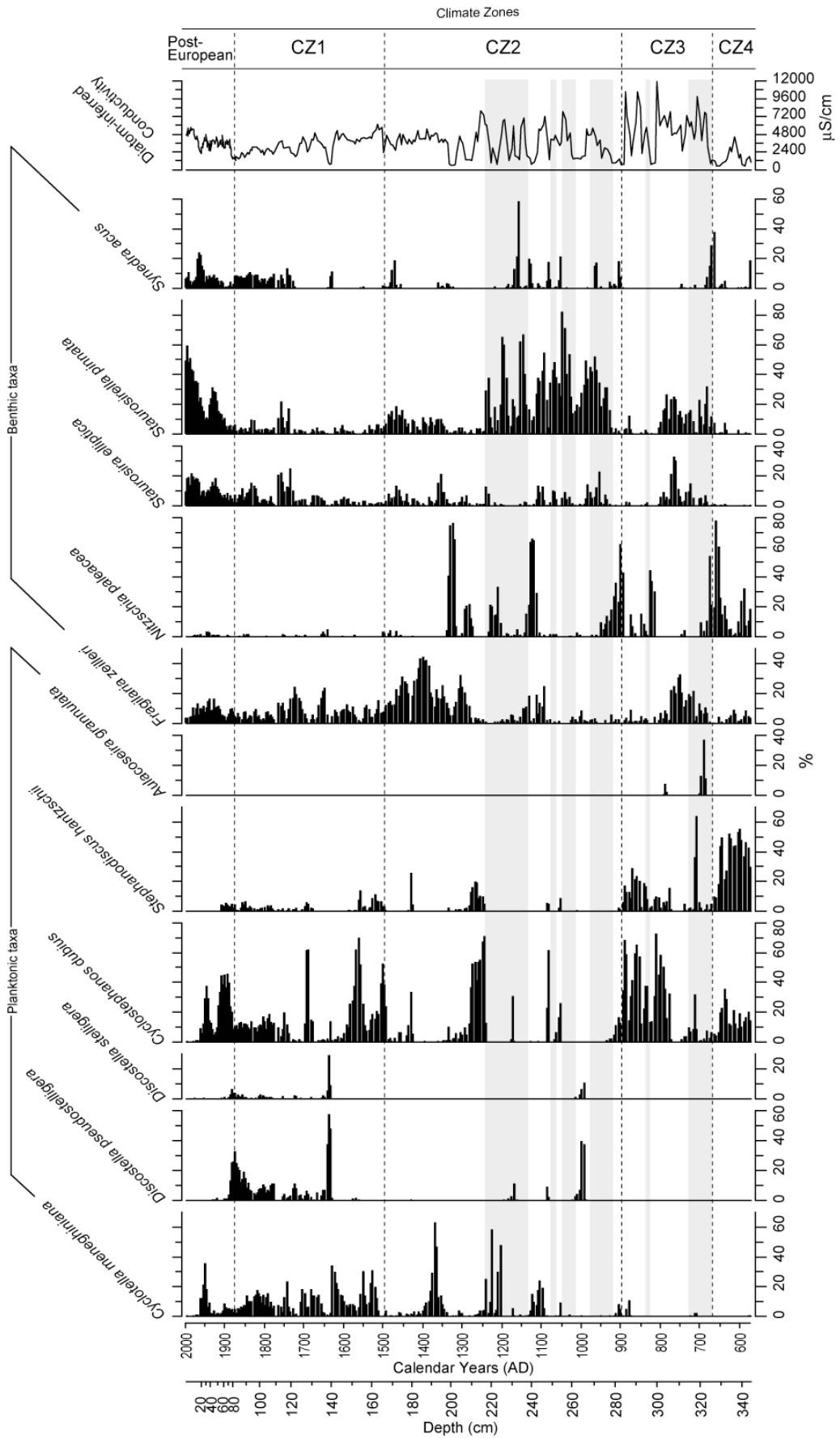


Figure 6.6: The full Lake Surprise fossil diatom record plotted by age. Only species that attained a relative abundance of 25% or more in at least one sample are included. Shaded areas indicate the presence of carbonate laminations in the sediment. Climate zones established in Section 6.5 are included for reference.

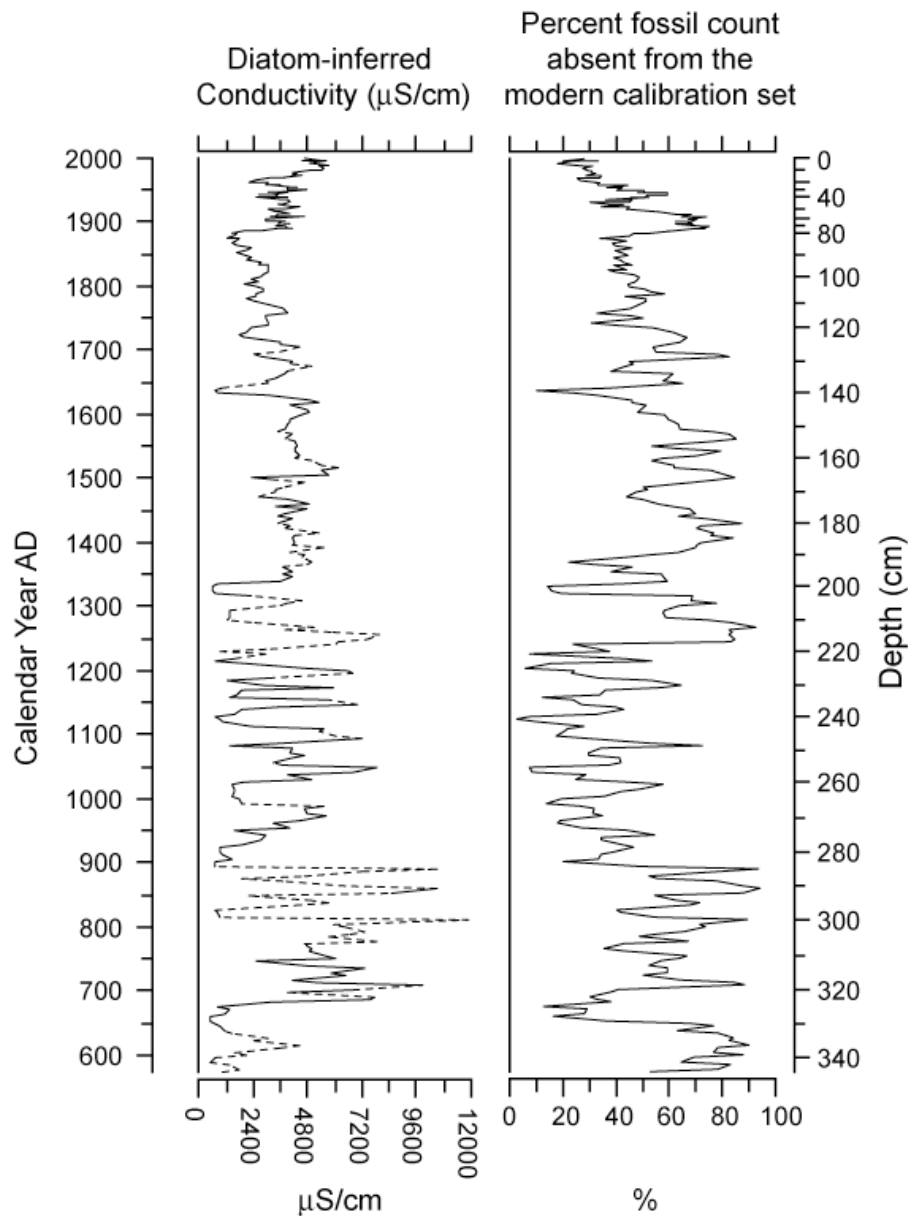


Figure 6.7: Diatom-inferred conductivity for the full Lake Surprise record and percent of fossil species absent from the modern calibration set. Dashed lines are reconstructions based on fossil diatom samples with poor modern analogues in the calibration set (see Section 5.6).

## 6.2 Comparisons of reconstructions with historical records

Birks (1998:313) states that “the most powerful means of validation is to compare the [palaeoenvironmental] reconstructions ... against known recorded historical environmental records.” This is an approach that has been widely applied to biota-based reconstructions (e.g., Fritz, 1990; Laird *et al.*, 1996a; 1996b, 1998; Cobb *et al.*, 2003; Calvo *et al.*, 2007) and its application to this study will be the focus of this section.

The underlying assumption behind palaeolimnological climate reconstructions is that a relationship exists “between the sediment record and the water column ... [and that this] can be calibrated with climate” (Battarbee, 2000:108). In relation to this study, the assumption is that fossil diatom assemblages preserved in the sediment column are a record of previous changes in the chemistry (specifically conductivity) of the water column. The hypothesis therefore, is that the fluctuations in the derived conductivity record reflect climatic changes. This hypothesis needs to be tested in order to determine the veracity of the results (Birks, 1998).

For the lakes to record shifts in moisture balance, they must be climatically sensitive, and there is an element of confidence that this is the case. Both sites were identified after the application of a strict set of criteria, which included the criterion of climatic sensitivity (Section 4.1.1). In the case of Lake Surprise, and in the absence of modern monitoring data, previous pollen and diatom analyses indicated a likely climate response at centennial and millennial scales (Figure 3.19), though annual to decadal-scale responsiveness remains undetermined. In the case of Lake Elingamite, this criterion was met due to observed lake level decline since the 1960s in conjunction with a hydrological model of the lake which indicated a clear climate response on annual timescales (Roger Jones unpublished data) (Section 3.3.4.1).

A comparison between the diatom-inferred (DI) conductivity and the ratio of benthic to planktonic diatom species in the Lake Elingamite record (Figure 6.8) provides further qualitative support for the notion that the lake is responsive to climatic changes. As discussed in Section 2.5, where suitable morphometry exists (Fritz, 2008), an increase in planktonic species can infer greater lake depth, while an increase in littoral/benthic species implies a shallower lake environment (Owen *et al.*, 1982; Gasse *et al.*, 1989; Barker *et al.*, 1994a; Brugam *et al.*, 1998; Tapia *et al.*, 2003). The morphometry of Lake Elingamite, with steep high walls and a broad flat bottom that would provide increased benthic and littoral habitat as lake-levels fell (Figure 3.13), lends itself to the application of such a qualitative measure. As illustrated in Figure 6.8, periods of increased DI conductivity from Lake Elingamite correspond well with a decreased occurrence of planktonic taxa, while periods of decreased DI conductivity align with an increase in percentage of planktonic taxa. This relationship is statistically significant with  $r^2 = 0.82$  ( $p < 0.0001$ ). When viewed in conjunction with the knowledge that the majority of the fossil diatom assemblages from this core have very good analogues in the modern

calibration set, this provides confidence that the DI conductivity reconstruction from Lake Elingamite, in the pre-European period at least, may be realistically recording fluctuations in effective moisture availability.

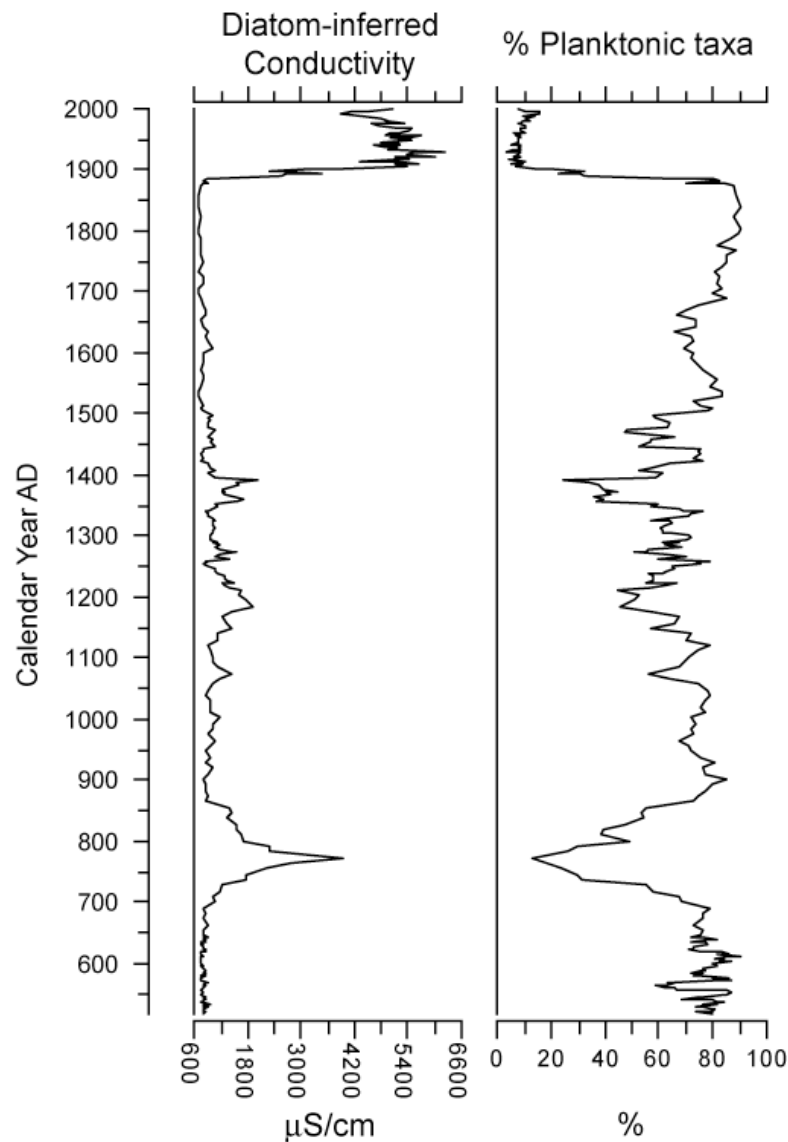


Figure 6.8: Diatom-inferred conductivity reconstruction and the percentage of planktonic taxa from the Lake Elingamite record.

However, despite the relationship between habitat and climate shown above, the record in the period that post-dates the impact of European arrival (*ca.* 1875 AD) does not reflect known climatic changes. A comparison of the DI conductivity since 1900, and the Palmer Drought Severity Index (PDSI, discussed further below) for the southeast of Australia (Ummenhofer *et al.*, 2009), is illustrated in Figure 6.9. Despite periods of both



positive and negative PDSI (in which negative values reflect drought conditions) the inferred conductivity record indicates very little overall change during the 20<sup>th</sup> century. After *ca.* 1875 AD, the diatom assemblages undergo a dramatic shift with the previously dominant *Discostella stelligera* and *D. pseudostelligera*, generally indicative of deep, clear lakes (Karst-Riddoch *et al.*, 2005; Moos *et al.*, 2005), being replaced by *Staurosirella pinnata*, *Staurosira construens* var. *construens* and *Cocconeis placentula*, which are generally associated with more turbid environments (Gell *et al.* 2002, 2005). While the exact cause of this shift towards turbid environments is not clear, it is plausible that it may be due to a combination of the construction and maintenance of a dirt road down into the crater, the grazing of livestock to the lake edge and the introduction of non-native fish to the lake. The recorded diatom assemblages are therefore more likely reflecting in-lake changes due to local landscape alteration and land use practices rather than climatic change. Furthermore, if a relationship could be established between the diatom data and recent climate, it would not be feasible to assume that this relationship is also present lower in the core where the diatom species assemblages are fundamentally different. For these reasons therefore, it is not feasible to calibrate the recent DI conductivity from Lake Elingamite against instrumental climate data.

NOTE:

This figure is included on page 208 of the print copy of the thesis held in the University of Adelaide Library.

*Figure 6.9: Diatom-inferred conductivity in Lake Elingamite since 1900 AD compared with 20th Century PDSI values for the southeast of Australia, calculated by Ummenhofer et al. (2009).*

In contrast to Lake Elingamite, Lake Surprise is located within the boundaries of Mount Eccles National Park and has been somewhat protected from disturbances, such as those mentioned above, making it the more likely of the two study sites to record recent climatic changes in the diatom data. While there is a dominance of benthic taxa toward the very top of the core, there is not a fundamental shift in the species composition, as was evident in the Lake Elingamite record. Indeed, the general species composition evident at the top of the core is replicated to various degrees lower in the record. This suggests that any disturbance to the lake that has occurred during the European period has not impacted the lake to such a degree as to fundamentally alter the diatom record, and that the driving mechanism behind the pre-European diatom assemblages is also likely to be driving the post-European assemblages. The assumption is that this driving mechanism is climate or, more specifically, fluctuations in the precipitation/evaporation (P/E) ratio, which has been identified as the “over-riding influence on lake levels [of the region]” (Jones *et al.*, 2001).

A physical representation of fluctuations in the P/E ratio is the occurrence of droughts and anomalous rainfall events. Anomalous rainfall events are easily calculated and are usually determined as deviations away from the World Meteorological Organisation standard baseline of 1951 – 1990 (Murphy and Timbal, 2008). Classifying droughts and drought intensity though, is more complex.

Droughts are usually classified through the derivation and application of various drought indices. Numerous such indices exist, each with individual benefits and limitations (Heim, 2002; Keyantash and Dracup, 2002; Mpelasoka *et al.*, 2008). The index used in this study is the Palmer Drought Severity Index (PDSI) which has the benefit of utilising localised data. It can therefore be applied to specific locations and, as the data is standardised, drought duration and intensity can be accurately compared between regions (Alley, 1984). The PDSI is calculated using an algorithm based on the supply and demand concept of water balance and takes into account rainfall, temperature and available water content of the soil (Palmer, 1965; Alley, 1984). The index generally ranges from -6 to 6, with negative values indicating drought conditions (Alley, 1984). Although the PDSI is widely used in the United States by government agencies such as the National Oceanic and Atmospheric Administration (see; [www.drought.noaa.gov](http://www.drought.noaa.gov)), it is not without its limitations. Some of the main criticisms centre on the short-term timing of the impacts of precipitation. The PDSI considers all

precipitation to be immediately available rainfall and, therefore, the time lag between snowfall and snowmelt, or the freeze and thaw of soils, means that the PDSI is not accurate in some regions on monthly to seasonal timescales (Alley, 1984; Dai *et al.* 2004). These criticisms are not applicable to this study though, due to the temporal resolution being examined and the fact that neither snowfall nor ice cover occur in the region.

The data presented here is from Ummenhofer *et al.* (2009) who have calculated the PDSI for the southeast of Australia for the 20<sup>th</sup> century. A comparison between the Lake Surprise DI conductivity, the PDSI, regional rainfall anomalies and effective precipitation (precipitation  $\div$  evaporation)<sup>4</sup> is presented in Figure 6.10. There is a clear and obvious relationship between the P/E ratio, rainfall anomalies and the PDSI. The severity of the current drought is also clearly demonstrated in an historical perspective and highlights the impact of rising temperatures on the nature of the recent drought in Australia (Nicholls, 2004).

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<sup>4</sup> The P/E data was presented in Figure 6.10 is derived using the calculated regional evaporation data of Jones *et al.* (2001) and Australian Bureau of Meteorology rainfall data. The data of Jones *et al.* (2001) is present only until 1990 and, while an attempt was made to extend this record to 2005 using linear estimation against the instrumental data of Jovanovic *et al.* (2008), the results were considered too inaccurate to present. Instrumental pan evaporation data of the region, taken from Jovanovic *et al.* (2008) (averaged from the only three sites in the study region), was compared to the data of Jones *et al.* (2001) for the overlapping period of 1970-1990. The poor correlation between the data ( $r^2 = 0.013$ ) meant that instrumental data could not be used to extend the record of Jones *et al.* (2001). Therefore, only the original evaporation data of Jones *et al.* (2001) was used and, while this only extends to 1990, adequately illustrates the relationship between effective precipitation and drought.

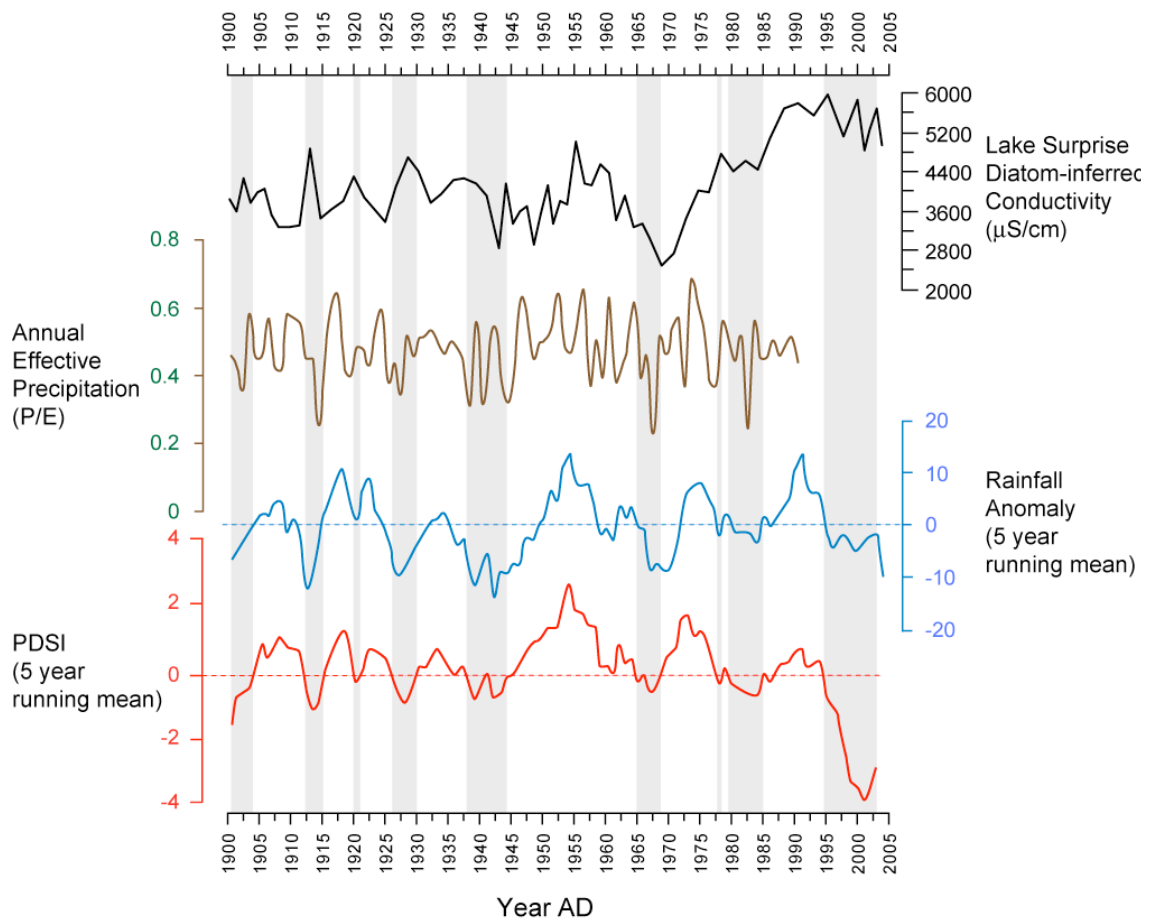


Figure 6.10: Comparisons between Lake Surprise DI conductivity and measures of regional rainfall and drought since 1900. Effective precipitation is derived by dividing annual regional rainfall data (Australian Bureau of Meteorology) by the evaporation data of Jones et al. (2001). Rainfall anomaly and PDSI from Ummenhofer et al. (2009). Shaded areas highlight periods of negative PDSI values.

The response of the lake to changes in moisture availability initially appears counter-intuitive, with periods of decreased moisture availability reflected by decreased conductivity and periods of increasing moisture availability reflected by increased conductivity, particularly since *ca.* 1935. Despite this, the DI conductivity record does show fluctuations similar to those of the other indices, albeit with the timing slightly skewed. The lack of relationship therefore appears to be more about chronological control than the efficacy of the transfer function and that the age of the reconstruction is offset. To examine this possibility, the DI conductivity has been compared to the PDSI, with the PDSI scale inverted to aid comparison (Figure 6.11).

In the early portion of the record, prior to *ca.* 1935, there is a good correspondence between the timing of peaks and troughs in the DI conductivity record and those in the

PDSI. This correspondence continues in the latter portion of the record, however, after *ca.* 1935, there is a pattern evident in which the DI conductivity changes consistently precede the changes in the PDSI reconstruction by approximately six years. The reasons for this apparent error are unclear. A discrepancy of six years falls well outside the statistical counting errors of the  $^{210}\text{Pb}$  dating, which range from  $\pm 0.5$  years to  $\pm 1.3$  years (Table 5.9), so this can be ruled out as a possible cause. There are, however, several other potential sources of error in age derivation.

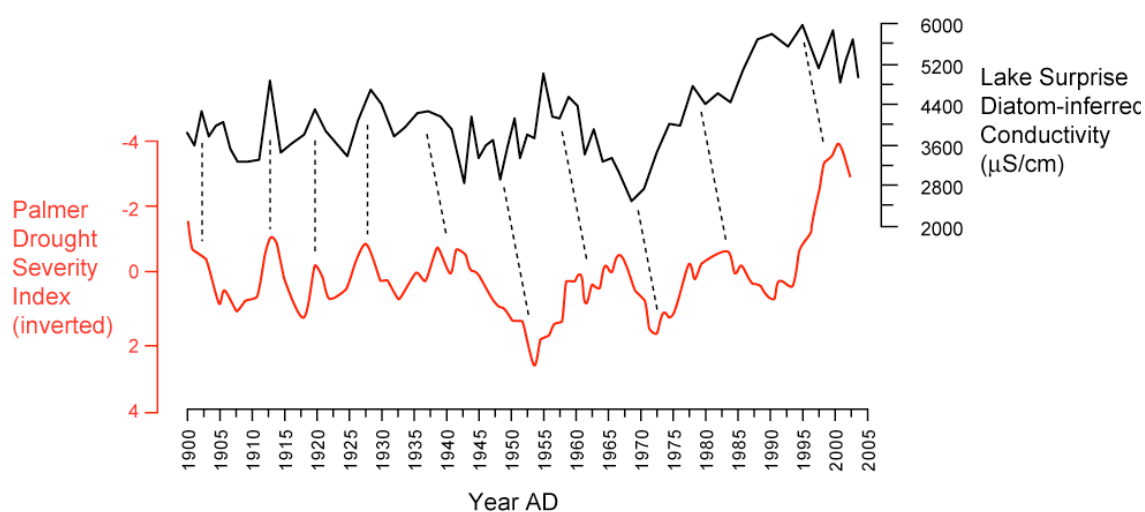


Figure 6.11: Comparison of Lake Surprise DI conductivity with the PDSI of Ummenhofer *et al.* (2009). Note that the PDSI scale is inverted for ease of comparison. Dashed lines indicate points of coherence.

Changes in the grain size of sediment have been found to affect the adsorption of unsupported  $^{210}\text{Pb}$ , with increasing surface area of sediments associated with an increase in unsupported  $^{210}\text{Pb}$  (He and Walling, 1996). However, in the absence of grain-size analysis, the relative impact of this can not be ascertained. A further potential cause of this error may lie in the occurrence of groundwater in the lake. The presence of  $^{222}\text{Rn}$ -rich groundwater interacting with the sediment can lead to an overestimation of  $^{210}\text{Po}$  and result in higher unsupported  $^{210}\text{Pb}$  calculations (Atun Zawadzki, pers. comm.; Ravichandran *et al.*, 1995). The occurrence of  $^{222}\text{Rn}$  in groundwater is common and the relative inputs of groundwater into lakes (Tuccimei *et al.*, 2004), streams (Cook *et al.*, 2003) or marine systems (Cable *et al.*, 1996; Schwartz, 2003), are often calculated through  $^{222}\text{Rn}$  measurement. The degree of  $^{222}\text{Rn}$  enrichment of the groundwater that feeds Lake Surprise is not known, however, sediment analysis from 46 – 1 cm of core

LSFS, which correspond to the portion of the core in which the ages are offset, indicate very high (min., 93%; max., 98.9%; avg., 96.7%; S.D., 1.3%) sediment moisture content (Figure 5.9). Again, it is unclear if this is the source of the dating error, though the fact that the remainder of core LSFS has an average moisture content of 95%, with more apparent dating accuracy, makes it unlikely.

Therefore, the cause, or causes, of the apparent dating error remain unclear. However, an examination of the  $^{210}\text{Pb}$  dating results from both study lakes demonstrates the diverging age determinations that can be obtained either by differing the method of measuring  $^{210}\text{Pb}$  decay (alpha- or gamma-spectrometry), or the method of calculating ages (CRS or CIC). Indeed, some degree of age offset is common in palaeolimnological studies. In a comparable study by Laird *et al.* (1998), a smoothed drought index showed the strongest correlation with diatom-inferred salinity when a three year time lag was factored in to the lake chronology. Similarly, a comparison of measured and DI salinity by Fritz (1990), shows an age offset of approximately ten years at times, though excellent correlation at others.

Therefore, while the cause of the age offset in Lake Surprise remains unknown, it is not an unexpected, or unprecedented, outcome. Moreover, when viewed in terms of the long-term record, an offset of around six years becomes less significant. More important is that a close relationship has been identified between diatom-inferred conductivity and the PDSI, which not only enables a long-term reconstruction of drought intensity, frequency and duration to be undertaken, it justifies the use of diatom-inferred conductivity as a proxy and provides further validation of the transfer function.

### **6.3 Lake climate records: Coherence and comparison**

Lakes respond to climatic forcing in different ways depending on factors such as landscape position (Webster *et al.*, 1996) and morphology (Mason *et al.*, 1994). Moreover, the degree of coherence between lakes responding to a common climate signal is affected by factors such as distance between the lakes, connectivity to groundwater (Magnuson *et al.*, 2004) and the climatic variable being examined (Benson *et al.*, 2000). In terms of the former, coherence can decrease with distance (Magnuson *et al.*, 2004), while in terms of the latter, site-specific climate responses, such as lake thermal variables, show weaker coherence than individual climate drivers (Benson *et*

*al.*, 2000). In this study, the distance between the two study sites is approximately 100 km, while the climate variable of interest is moisture availability or, more specifically, the PDSI.

The PDSI used in the comparisons above is a regional climate index reflecting moisture availability for south-eastern Australia (Ummenhofer *et al.*, 2009) and the Lake Surprise DI conductivity record can therefore be interpreted as reflecting regional climate. However, due to reasons outlined above, the Lake Elingamite record is not amenable to such comparisons in the post-European phase and, while it is likely that the lake is responding to a regional scale climate signal, such as that proposed by Jones *et al.* (2001), this is yet to be determined. A comparison between the two complete lake records will clarify this. A coherent response of the two lakes over the extended time frame would confirm that Lake Elingamite is also responding to the same climate signal, and that the DI conductivity reconstruction from Lake Elingamite is not reflecting a site-specific signal (*sensu* Fritz, 2008).

Figure 6.12 shows a comparison between the DI conductivity of both lakes over the full length of their records. While it is immediately apparent that the Lake Surprise record demonstrates far greater absolute variability than that of Lake Elingamite, there is also a clear coherence between the records with five notable time periods evident which occur synchronously. These have been labelled as the post-European phase and ‘climatic zones’ one to four (CZ1 - CZ4) and are the focus of the discussion in Section 6.5 below, in which the variability and trends within these zones is also discussed.

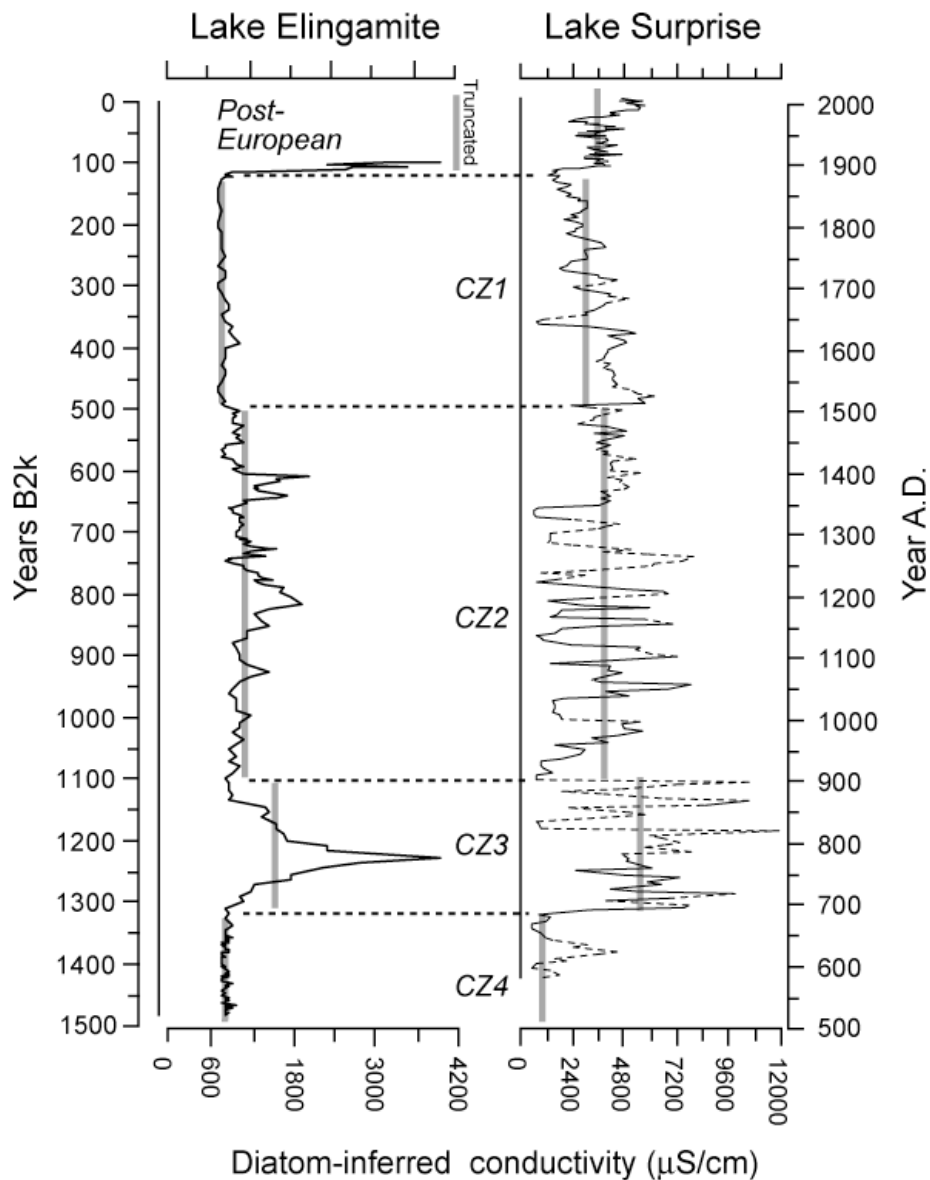


Figure 6.12: Comparisons between the long-term record of DI conductivity from Lakes Elingamite (truncated in the post-European phase) and Surprise showing a coherence between the lakes' signals. Grey bars indicate the average DI conductivity of the climate zones, calculated only from samples with good modern analogues (see Section 5.6). Dashed lines represent portions of the reconstruction with poor analogues in the modern calibration set.

In the Lake Elingamite record, climatic zones one and four are remarkably similar, with both being characterised by lengthy periods of low conductivity, averaging 800 and 816  $\mu\text{S}/\text{cm}$  respectively. Such persistently low values suggest that this is the lower limit of conductivity for this lake and that the lake was probably full. These periods initially appear to be climatically stable, yet comparisons with the Lake Surprise record indicate that notable climatic variability was still present during these times. The Lake Surprise record, during CZ1, shows decreasing conductivity, yet no such trend is apparent from



Lake Elingamite, in which conductivity values remain relatively constant. This suggests that the lake may have reached its capacity during this period and may have been overflowing, a situation observed in other western Victorian lakes at the time of European settlement of the region (Currey, 1970). The similar low conductivity values and degree of stability in Lake Elingamite during CZ4, suggest that the same may have been occurring at that time as well.

Some support for this hypothesis is evident in comparisons between the fossil diatom composition and the modern calibration set. The diatom record from Lake Elingamite, during zones CZ1 and CZ4, are dominated by *Discostella stelligera* and *D. pseudostelligera*, which together account for between 60% and 90% of the diatom assemblages during these phases. By way of comparison, these species also dominate the modern diatom assemblage of Blue Lake, which is a deep (~77 m), oligotrophic crater lake (Lamontagne and Herczeg, 2002) with a measured conductivity of 636  $\mu\text{S}/\text{cm}$ . In the surface sediment sample from this lake, *D. stelligera* and *D. pseudostelligera* account for 83% of all valves counted. This is the only site in the modern calibration set in which these species dominate the assemblage to such a degree. Studies from elsewhere though, have also noted these taxa have a preference for deep, chemically dilute lake environments (Catalan *et al.*, 2002, Moos *et al.*, 2005; Laird and Cumming, 2008), thus providing some support to the theory that, not only was Lake Elingamite fresh during CZ1 and CZ4, it was also deep.

Zone CZ3 is a period of elevated conductivity in Lake Elingamite which is also present in the Lake Surprise record, albeit with a far greater degree of variability. Slightly lower amplitude of variability, and generally lower conductivity values, are present in zone CZ2 in Lake Surprise, which, again, is a state that is replicated in the Lake Elingamite record.

The coherent nature of the DI conductivity signals in both lakes, combined with the established relationship between Lake Surprise DI conductivity and moisture availability in the historical record, demonstrates that both lakes are recording a biotic response to a common regional climate signal. However, it is apparent that, while both lakes respond to the low amplitude trends, Lake Surprise appears more responsive to high-amplitude climatic variability. The presence of two differing, but synchronous, records provides an added dimension to the interpretation of the regional climate

history. Lake Elingamite is less sensitive than Lake Surprise in terms of a biotic response to climatic forcing, yet the very good analogues in the modern calibration set instill confidence in the derived conductivity record. This, in turn, supports the interpretation of the Lake Surprise DI conductivity record during the portions of that record which have poor modern analogues. The greater sensitivity of Lake Surprise provides an insight into the degree of variability within the climate record which would not be evident if examining Lake Elingamite in isolation. The combination of the two records therefore, enables a comprehensive regional-scale reconstruction of moisture shifts over the last *ca.* 1500 years.

#### **6.4 Interpretative model for the lake records**

As discussed in Section 2.5, the utility of diatoms makes them a powerful tool for reconstructing past lake conditions, yet there are several issues which can confound the interpretation of diatom-inferred results. For example, Gasse *et al.* (1997:547) implicitly warn against drawing linear relationships between diatom-inferred lake salinity and climatic change, citing several aspects of local hydrology which “may modify the relationship between salinity change and climatic forcing.” Similarly, Fritz (2008) cautions that morphological features can effect the nature of a lake’s response to climate forcing, while Saros and Fritz (2000) suggest that changes in diatom species composition may, at times, be driven by interactions between ionic composition and/or concentration and their influence on nutrient dynamics. Ideally, these issues would be addressed, and the interpretation strengthened, by the addition of further proxies and/or by comparisons against results of separate climate studies from the same region. Where these options are unavailable though, developing a diatom inference model tailored specifically for reconstructing the environmental variable of interest from a specific location can still prove effective (e.g., Ryves *et al.*, 2002). However, the limitations of a single-proxy approach should be recognised and interpretation of the resulting data undertaken with caution.

In recognition of this, interpretive models will be developed for each study site, proposing the manner in which the lakes respond to climate change. These are developed by considering all the evidence available, and will provide a framework for the interpretation of the climate record that will follow in Section 6.5. This approach is undertaken in recognition of the fact that, at times in the records, and despite the best

efforts in project design, there are insufficient data to fully understand the complexities of the lake-climate relationship. In addition, this approach also takes in to account other variables, which are not reflected in the conductivity record, such as patterns of species turnover for instance, as these may also contribute to an understanding of changing palaeoclimates of the region.

#### 6.4.1 Lake Surprise

Investigations from Lake Surprise show notable variability both in terms of the fossil diatom record and in regard to the history of the lake itself. In terms of the fossil diatom records, Lake Surprise is the most complex of the two study sites, with large and rapid shifts in species composition evident. However, the interpretation of this record is hampered by several factors, including poor modern analogues in parts of the fossil diatom record and the absence of ionic chemistry from the modern calibration set (Section 4.3.1 explains the reasons behind this). Sedimentary analysis also indicates marked variability within the moisture, organic and carbonate content of the sediment, suggesting considerable in-lake changes through time. Though again, the interpretation of these is made more difficult by the lack of a detailed hydrological model of the lake, while project constraints precluded geochemical and/or isotopic analysis of the sediment, which would also have been instructive.

However, despite the absence of some instructive analyses, the available data can prove informative to the climate reconstruction. Sediment analysis has long been employed for reconstructing climate change (e.g. Bowler, 1981; Last and De Deckker, 1990; Mooney, 1997) as the results can be considered to be a reflection of the relationship between the lake and climate. The following discussion will examine the available sediment data and propose likely climate mechanisms responsible for the identified changes. The second part of this discussion examines the evidence of species turnover in the fossil diatom record, with specific reference to the resource requirements of the key planktonic species *Cyclostephanos dubius* and *Stephanodiscus hantzschii*. As these species are absent from the modern calibration set, they do not contribute to the conductivity reconstruction. However, the dramatic fluctuations in the relative abundance of these species reflect in-lake changes that need to be evaluated in order to ascertain their contribution to the climate reconstruction.

Figure 6.13 illustrates the sedimentary and DI conductivity data of the Lake Surprise record. Of primary interest is the presence of carbonate-rich laminations<sup>5</sup> in the sediment, and the corresponding variability in the DI conductivity and sedimentary carbonate and organic content profiles. In regard to the sedimentary carbonate content, there are three key processes which require further discussion. These are; the source of the carbonate in the lake, the factors that lead to carbonate precipitation from the water column and the factors that lead to the preservation of laminations in some periods and the lack of formation and/or preservation in others.

The most logical source of carbonate is the limestone which underlies the volcanic plains (see Section 3.1), and the regional water table which carries the carbonate to the lake. Timms (1975) found the cation dominance of the lake water, sampled in 1971 when the PDSI was only slightly negative (Figure 6.10), to be  $\text{Na} > \text{Mg} > \text{Ca} > \text{K}$ . However water samples for this project, taken in 2004 when the PDSI had been strongly negative for nine years, showed a dominance cascade of  $\text{Na} > \text{Ca} > \text{Mg} > \text{K}$ . The implication here is that, as the lake evaporates during dry phases, the relative proportion of lake volume contributed by groundwater sources increases, thereby increasing the importance of Ca in the cation dominance cascade.

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<sup>5</sup> Cohen (2003) defines laminated sediments as “Sedimentary deposits consisting of extremely thin layers of (generally very fine-grained) sediment” (p.402) as opposed to ‘varves’ which are an “annual set of layers ... in sediments” (p.404). Within the laminated sediments there are fine carbonate layers followed by fine organic layers. While it is possible that these may represent seasonal or annual layers of deposition, it is not possible to test this with available dating tools. Therefore, the term ‘laminations’ is more appropriate.

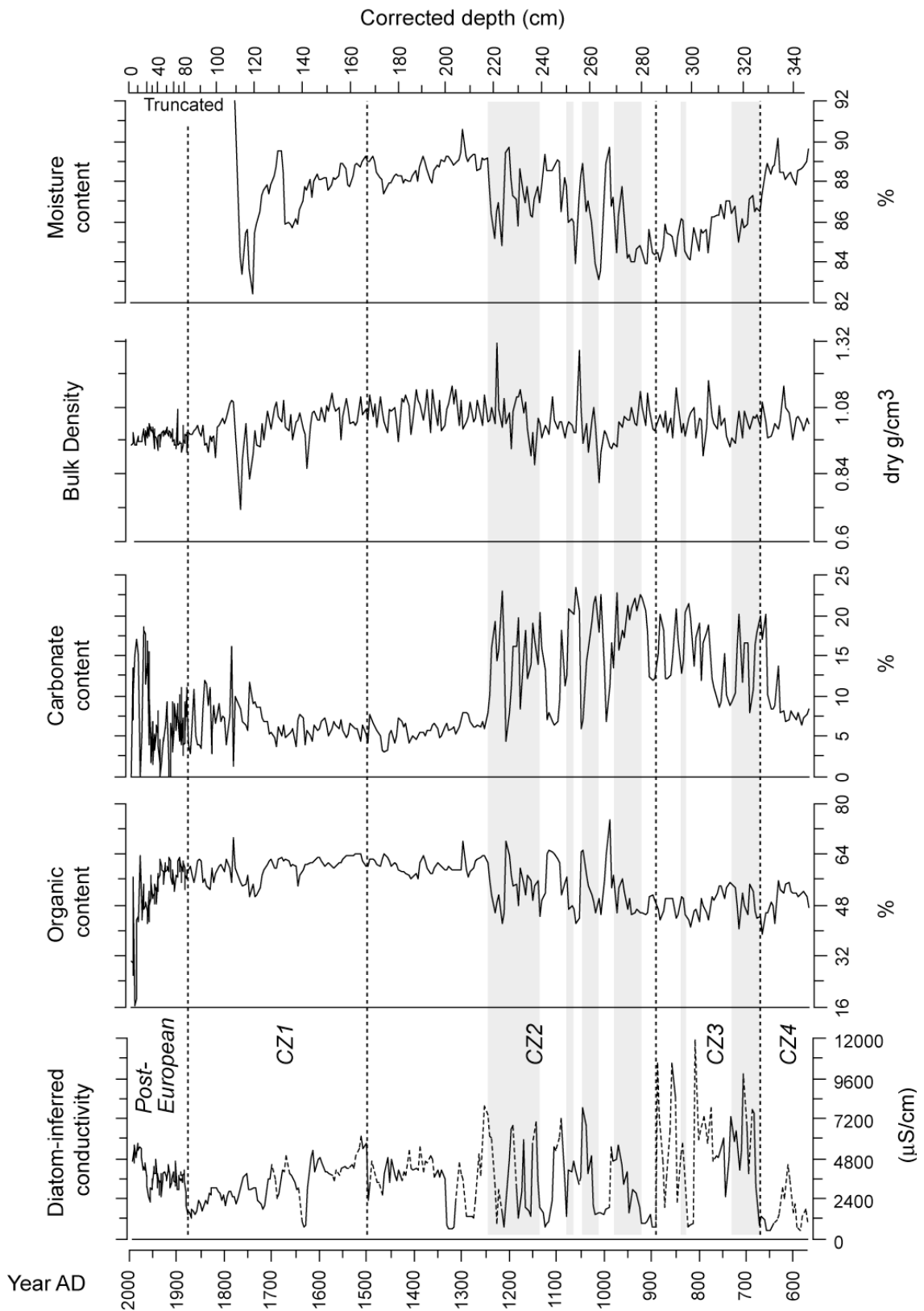


Figure 6.13: Summary diagram of DI conductivity and results of sedimentary analyses from the full Lake Surprise record. Shaded areas represent periods where carbonate laminations are evident in the sediment. Dashed lines in the DI conductivity graph represent poor analogues in the modern calibration set. Moisture content graph is truncated for ease of illustration.

There is some support for this assertion from both the historical record and the fossil diatom records. Due to the neutralizing effects of carbonate, an increasing prevalence of groundwater should be reflected by an increasing pH. This was the case when the lake was studied by Timms (1975), who found that pH increased from 7.7 and 7.8 in July and November respectively, to 8.5 by the end of summer in February. Sampling for this project in early March 2004 identified a pH of 8.75. In addition, the fossil diatom assemblage during the periods where laminations are present contain the species *Encyonopsis ruttneri*, *E. microcephala*, *Gomphonema affine* and *Achnantheidium minutissimum*, which are rarely evident, or completely absent, at other times in the record (Figures 5.24 and 5.25). While the ecological preferences of *E. ruttneri* are unknown, the other three taxa have a preference for alkaline to circum-neutral waters (Kelly, 2000; Sonneman *et al.*, 2000). In addition, *Cyclostephanos dubius* has been found to be inhibited by waters of high alkalinity (Clarke, 1989) and this species is absent, or its prevalence greatly reduced, during periods where laminations exist.

However, it is recognised that there may be complex interactions between numerous factors that could be influencing lake water pH, with fluctuating groundwater dominance in the hydrological budget of the lake being just one of these factors. In the absence of long-term lake monitoring data, it is difficult to speculate on the relative contribution of the various potential drivers of pH change. Based on the limited data available from Timms (1975) and this project though, it appears that, at seasonal scales, there is a pH increase in the warmer months, while at sub-decadal scales, recent decades have seen and increase in sedimentary carbonate content (Figure 6.13 above).

The measurement of ionic chemistry in the routine analysis of the full suite of water samples used in this study may have greatly aided the interpretation of this record. Diatoms are highly sensitive to changes in ionic chemistry (Gasse *et al.*, 1997) and ionic ratios often explain a significant amount of variation in the distribution of diatom species in transfer functions (e.g., Gell, 1997; Ryves *et al.*, 2002). The ability to reconstruct ionic parameters from the fossil diatom data may have provided more data to support and/or guide this interpretation.

Nonetheless, the factors discussed above support the assumption that the aquifer which maintains the lake is the primary source of carbonates. By extension, this implies that periods of the record where sedimentary carbonate content is elevated equate to periods

where groundwater is relatively more important in the hydrological budget. However, if this were the case, then it is reasonable to expect that the opposite is also true. That is, in times where precipitation is more dominant in the lake's hydrological budget, sedimentary carbonate content would be low.

In support of this, similar results were found in the Holocene records from East and West Basin, two maar lakes which lie within the study region (Last and DeDeckker, 1990). In this instance, an increasingly humid climate "about 9,000 yr BP" was reflected in low total carbonate content in the sediments. The authors argue that the sedimentology at this time was "probably the result of a change in the hydrological budget of the lake: an increase in the amount (or proportion) of surface runoff ... to the lake, relative to the earlier dominance of groundwater influx" (Last and DeDeckker, 1990:979). Similarly, increased aridity between 5,000 and 3,000 years BP is reflected by more saline conditions in the lakes which is "marked by an increase in total carbonate content" (Last and DeDeckker, 1990:979).

Other studies from crater lakes in western Victoria have also invoked warmer climates as the cause of increased carbonate deposition. At Lake Keilamete for example, Bowler (1981) interprets elevated carbonate levels at *ca.* 5500 BP as being indicative of a shift towards negative water balance. Bowler (1981:437) also identified higher carbonate content between *ca.* 600 – 700 BP and proposes that this may be "reflecting a return to slightly warmer or drier environments" though, from the same lake, Mooney (1997:147) suggests that a decrease in organic carbon at the time may indicate that "warmth was not a factor [in this change]." From Lake Wangoom, a maar in the south of the region, Edney *et al.* (1990) reconstruct late Pleistocene and Holocene climates using pollen, microfaunal and sedimentological evidence. From approximately 22,000 to 9,000 years BP, they interpret high inorganic and carbonate content as indicative of "extremely dry conditions" (Edney *et al.*, 1990:337), which is supported by a lack of pollen in the core during this time. Similarly, from *ca.* 7,000 years BP, low carbonate and high organic content is interpreted as being indicative of the maintenance of a deep water lake.

Similar conditions of low carbonate and high organic content are evident in the Lake Surprise record between *ca.* 1250 – 1750 AD. In this section of the record, the organic content of the sediment is high, carbonate content is consistently low and all the measured sediment variables show a degree of stability that is absent from the

remainder of the record. Over this ~500 year period, the inferred conductivity shows evidence of freshening lake waters. This is mirrored in the Lake Elingamite record which suggests 'lake full' conditions during this time. The indication from both records therefore, suggests that a positive moisture balance existed during this time, with increased effective precipitation resulting in a stable, deep lake, leading to organic-rich sediment deposition which is likely to be mainly autochthonous in nature. In light of the interpretations of other sedimentary sequences outlined above, this set of conditions is interpreted as reflecting an environment in which, relative to other ions, there is less carbonate in the lake water due to the hydrological budget being dominated by precipitation, rather than groundwater influence.

The above provides the first tenet of the interpretive framework. That is, that a relationship exists between increased sedimentary carbonate content and decreased effective precipitation. The second process that requires elaboration is the precipitation of carbonate from the lake water and the factors that may facilitate precipitation.

Lake water can remain saturated with carbonate until carbonate precipitation occurs due to endogenic or authigenic factors. At present, and in the absence of knowledge of the specific geochemical components of the carbonates or the photosynthetic rate of the lake for example, the precise origin of precipitated carbonates remains unknown. One potential mechanism though, may be organic productivity. Events such as algal blooms have been found to facilitate carbonate precipitation through the provision of nucleating surfaces (Thompson *et al.*, 1997; Hodell *et al.*, 1998), and this is a possible mechanism in Lake Surprise which is naturally eutrophic (Timms, 1975) and presently supports blooms of green and blue-green algae (Paul Leahy, pers. comm.).

A comparison of the diatom and carbonate records indicates that peaks in the number of diatom valves per dry gram occur prior to periods of lamination. These peaks are primarily driven by blooms of *C. dubius* and *S. hantzschii* which occur prior to most laminations, before dissipating during the lamination periods. These species co-occur through the majority of the record and this, combined with observations from the literature, suggests that they have overlapping ecological requirements. Both species have an affinity for high phosphorus (Kilham *et al.*, 1986; Anderson, 1990; Bradshaw and Anderson, 2003), and *S. hantzschii* has been identified as being in the low Si:P category of resource relationships, as proposed by Kilham *et al.* (1986). Their co-



occurrence in the record suggests that *C. dubius* may also fall in to this category. Kilham *et al.* (1986) propose that blooms of low Si:P species occur until the supply of available silica is exhausted, which in turn leads to blooms in forms of algae, such as green and blue-green algae, that do not require silica. The increased photosynthetic activity that occurs during such blooms reduces CO<sub>2</sub> levels (Cohen, 2003) which, in conjunction with the availability of nucleating surfaces, leads to a situation conducive to the precipitation of carbonates (Thompson *et al.*, 1997; Hodell *et al.*, 1998; Cohen, 2003).

However, while organic production (i.e., ‘algal’ blooms) represents a potential mechanism for carbonate precipitation in Lake Surprise, this can not be conclusively determined. Hence, any interpretation of the climatic significance of lamination formation must be undertaken with caution.

The third point that requires elaboration is the process, or processes, that enable the preservation of carbonate, and carbonate laminations, within the sediment. In the Lake Surprise record, there are three distinct periods where sedimentary carbonate content is elevated. The first, from 327 cm to 280 cm (668 – 981 AD), shows a trend of increasing carbonates overlying high frequency variability with few laminations. The second, from 280 cm to 218 cm (981 – 1244 AD), shows no discernable long-term trend, has a lower frequency, though greater amplitude, of variability than the previous period and each peak in the carbonate record corresponds to a period of laminations. The final period is from the mid-1960s to the present. No laminations are evident in during this portion of the record, possibly due to the high moisture content and unconsolidated nature of the sediment, and it is therefore difficult to determine if this 45 year period is equivalent to either of the previous episodes of carbonate deposition. Given that this period is too short to identify any real trends or make meaningful comparisons, it will not be discussed in detail in this section.

In the earliest period, the inferred conductivity from both lakes is elevated, supporting the notion of decreased effective moisture and an increased dominance of groundwater (relative to rainfall) in the hydrological budget of the lake. Despite the presence of numerous peaks in the carbonate record though, there are only two sections where carbonate laminations exist, from 327 to 317 cm (668 – 722 AD) and from 298 to 296 cm (835 – 836 AD). Therefore, a situation exists in which carbonate is being

precipitated and deposited in the sediment, though rarely in the form of laminations. Where the laminations do exist though, they occur as the inferred conductivity is increasing and, by extension, as the lake was becoming shallower and groundwater more important in the hydrological budget.

The second period of carbonate deposition, from 280 cm to 218 cm (981 – 1244 AD), shows a lower frequency, though greater amplitude, of variation than the preceding phase. There are four sections of elevated sedimentary carbonate, each lasting longer than those in the previous period and each being associated with preserved laminations. In further contrast to the earlier period, inferred conductivity is generally low, though this increases quickly in line with the rapid increases of the carbonate record. In addition, the periods of low conductivity correspond to times of high organic content in the sediments, indicative of a productive, deep lake environment (Cohen, 2003).

There are a number of elements required to facilitate the preservation of carbonate laminations, though foremost amongst these is the pH of the lake water. Carbonate will dissolve under acidic conditions and therefore, alkaline conditions are essential for its preservation (Cohen, 2003). However, the increased primary productivity which can facilitate carbonate precipitation can also lead to its dissolution during lake stratification. As the algal remains decompose, these can release CO<sub>2</sub>, thereby decreasing the pH of the hypolimnion through the formation of carbonic acid, resulting in the dissolution of carbonate in the sediment (Dean, 1999; Cohen, 2003). This is especially the case where an anoxic hypolimnion is present (Dean, 1999). In contrast, the existence of an anoxic hypolimnion can aid in the preservation of laminations because it deters bioturbation (Wehrli *et al.*, 1997; Teranes *et al.*, 1999, Dean *et al.*, 2002). Furthermore, the dissolution of a small percentage of calcite in the hypolimnion can lead to equilibrium in the lake waters (Gruber *et al.*, 2000), thereby aiding the preservation of laminations by preventing further dissolution. Finally, if laminations are to be preserved, there is a need for the surface sediments to be protected from turbulent mixing of the surface sediment layers.

To summarise, if laminations are to be preserved, there is a requirement for an alkaline to circum-neutral anoxic hypolimnion which is protected from turbulence. In the case of Lake Surprise, wind-driven turbulence is highly unlikely as the lake is well protected by the high crater walls (Figure 3.15). The processes that would lead to alkaline or circum-

neutral conditions in the hypolimnion can not be determined at this stage, however, the most likely mechanism that would cause the deoxygenation of the hypolimnion is lake stratification. This would also act as a deterrent to bioturbation. The fact that the lake currently has high carbonate levels and stratifies in summer without preserving laminations, suggests that a longer period of stratification may be required for laminations to be deposited and remain undisturbed. This may have taken the form of increased temperatures during spring and autumn, though the possibility of meromixis can not be discounted.

In much of the record, there are numerous peaks in the relative abundance of the planktonic taxa *Cyclostephanos dubius*, *Stephanodiscus hantzschii* and *Cyclotella meneghiniana*, with corresponding rapid transitions between benthic-dominated and planktonic-dominated species assemblages (Figures 5.31 and 5.32). The fine-resolution sampling regime enables observations of diatom species turnover to be made and, while these will be discussed further in Section 6.5 in relation to specific occurrences in the record, some general observations can be made from the longer term record which may provide insight into the mechanisms driving the species shifts.

As noted above, planktonic diatoms with low Si:P resource requirements (Kilham *et al.*, 1986), namely *Cyclostephanos dubius* and *Stephanodiscus hantzschii*, occur frequently throughout the record and often dominate the species assemblage. Given the small catchment size, the most likely source of phosphorus in Lake Surprise is nutrient recirculation (Cohen, 2003). This indicates that increases in the relative abundance of these two species most likely reflect a well mixed lake environment. Consistent with such a hypothesis, it is interesting to note that these species rarely occur during periods of carbonate lamination when, it is proposed, the lake stratifies for longer periods. However, soon after the periods of lamination cease, these species often return to dominate the record. Tracey *et al.* (1996) found that, when meromixis was sufficiently developed so as to significantly reduce phosphorus flux from the sediments, diatom flora shifted towards an assemblage indicative of lower trophic status. While evidence of meromixis can not be definitively derived from the Lake Surprise record, an extended period of stratification is proposed as a mechanism facilitating the formation and preservation of the carbonate laminations. The shifts in diatom species assemblages before, during and after periods of lamination, appear to be support this proposal. After laminations cease, a return of a species assemblage indicative of an increased trophic

state, such as the *C. dubius* and *S. hantzschii* dominated assemblages seen here, suggests that phosphorus flux from the sediments may increase once a mixing regime is reestablished.

Neither *C. dubius* or *S. hantzschii* were identified in the modern training-set lakes and, therefore, the periods where they dominate the fossil diatom assemblages have poor modern analogues. As a result, confidence in the conductivity reconstructions of these portions of the record is low, though typically, these species are considered to be freshwater taxa. It is interesting to note that these species are rarely found in modern diatom assemblages in Australia, being completely absent from the salinity transfer functions of Gell (1997) and Taukulis and John (2009). *Cyclostephanos dubius* was only occasionally present in low abundances in the TP transfer function of Tibby (2004), while *S. hantzschii* was not identified. Similarly, Tibby and Reid (2004), found *C. dubius*, though not *S. hantzschii*, in the modern training set of a diatom-salinity transfer function from planktonic taxa in the Murray River. The authors determined a conductivity optimum of 620  $\mu\text{S}/\text{cm}$  and a tolerance of 215  $\mu\text{S}/\text{cm}$  for *C. dubius*.

The implications of the scarcity of these species in contemporary Australian lake settings are two-fold. Firstly, it indicates that no contemporary analogues of many portions of the Lake Surprise record have been identified in Australia. This not only highlights the disparity between past and present lake conditions, it makes inferences about past lake conditions, when these species dominate the fossil assemblages, difficult. Secondly, though related to the first point, it implies that information regarding the ecology of these species must be retrieved from further afield. Often, this is only available from datasets in which the environmental gradient for salinity/conductivity is very short and often biased toward low salinity sites, and in which salinity/conductivity is not identified as a key driver of species abundance (e.g., Bennion, 1994; Tibby, 2004). Indeed, these species are most commonly identified in nutrient-transfer functions and only rarely found in statistically informative abundances in salinity-transfer functions. Hence, both species are recognised more for their status as nutrient indicators than salinity indicators (e.g., Kilham *et al.*, 1996; Bradshaw and Anderson, 2001), and as such, their appearances in fossil records are commonly interpreted as indicative of eutrophic conditions (e.g., Håkansson and Regnéll, 1993; Bennion, 1995; Sayer, 2001).

Based on these observations, the implication is that nutrient change, rather than salinity change, may exert a greater influence on the occurrence of these species. However, it may not be a simple case of being one or the other, as interactions between these two variables should also be considered. Saros and Fritz (2000) argue that salinity and ionic composition influence nutrient dynamics in saline lakes and these interactions may, in turn, influence shifts in diatom species composition. Furthermore, the authors argue that the high end of a taxon's salinity tolerance may be expanded by nutrient enrichment, citing the example of Fritz *et al.* (1993) who noted the occurrence of *C. dubius* and *S. hantzschii* in eutrophic hyposaline waters. In this regard, and despite the fact that conductivity was identified as the variable exerting the greatest influence over the modern diatom assemblages (see Section 4.3.3), the relative influence of conductivity may be diminished at times in the Lake Surprise record where *C. dubius* and *S. hantzschii* dominate the species assemblage. In light of this, and considering the poorly defined conductivity requirements of these species, it is prudent to interpret their relative abundances in the Lake Surprise record more in terms of their nutrient requirements, and how this may inform the climate history, than in terms of conductivity.

In light of the above discussion, the following interpretive model is developed, which provides the basis of the climate reconstruction presented in Section 6.5 below. Sections of the core in which sediment carbonate content is low and organic content is high, are considered representative of deep-lake conditions with high precipitation. This is supported by the fossil record in which periods of increased organic content of the sediment generally correspond to decreased inferred conductivity. Conversely, sections of the core with elevated sedimentary carbonate content are assumed to be indicative of a negative moisture balance and an increased relative importance of groundwater in the hydrological budget of the lake. Where these periods correspond with the preservation of carbonate as laminations, an extended period of lake stratification is invoked as the most likely cause. At this stage though, there is insufficient data to propose a possible cause of elevated pH in the hypolimnion or identify the mechanisms that drive carbonate precipitation in the lake. In terms of the diatom record, periods in which *Cyclostephanos dubius* and *Stephanodiscus hantzschii* dominate the record are considered to be indicative of a well-mixed lake in which nutrient recirculation is occurring.

#### 6.4.2 Lake Elingamite

Of particular interest in the Lake Elingamite record is the gradual shift within the dominant species of *Discostella stelligera* and *D. pseudostelligera* throughout the pre-European portion of the record. Prior to *ca.* 850 AD, *D. stelligera* dominates the record, however after this time, its relative abundance gradually decreases in line with a concurrent increase in *D. pseudostelligera*. This suggests that the lake may be recording subtle changes in the climate of the time and as such, further examination of these species is warranted.

Under light microscopy, these species can be difficult to separate (Kling and Håkansson, 1988; Sonneman *et al.*, 2000) and Haworth and Howley (1984) even propose that *D. pseudostelligera* is merely a variety of *D. stelligera*. In this study, the species were separated in the manner proposed by Krammer and Lange-Bertalot (1991) on the basis of the size of the cell, with *D. pseudostelligera* being  $< 10 \mu\text{m}$  and *D. stelligera* being  $> 10 \mu\text{m}$ , and on the presence (*D. pseudostelligera*) or absence (*D. stelligera*) of bifurcated striae (Kling and Håkansson, 1988). When these species co-occur in palaeolimnological investigations, their relative abundances are occasionally combined in the final analysis and referred to as a ‘complex’ (e.g., Rühland *et al.*, 2003; Tapia *et al.*, 2003), however, the variation within the abundances of these taxa in the Lake Elingamite record implies that there is some value in keeping these species separate.

The conductivity optima of *D. stelligera* and *D. pseudostelligera* were both very fresh at  $883 \mu\text{S/cm}$  and  $783 \mu\text{S/cm}$  respectively, though lower optima ( $127 \mu\text{S/cm}$  and  $195 \mu\text{S/cm}$  respectively) were derived in a transfer function developed from Murray River (Australia) samples (Tibby and Reid, 2004). Despite being found in some riverine environments (e.g., Tibby and Reid, 2004), the species are more commonly found in lacustrine settings (Sonneman *et al.*, 2000). The species prefer waters greater than eight metres deep (Catalan *et al.*, 2002; Moos *et al.*, 2005; Laird and Cumming, 2008) and are commonly found in clear, chemically dilute, oligotrophic lakes (Karst-Riddoch *et al.*, 2005; Reid, 2005). In Australia, *D. stelligera* was found to dominate the pre-European diatom assemblage in billabongs along the Yarra River (Victoria) before rapidly declining following European landscape disturbance (Leahy *et al.*, 2005; Leahy, 2007), suggesting that turbidity may be a factor in limiting the species’ abundance.

In modern calibration-sets, little separates these species in terms of pH preferences, as both species consistently demonstrate a preference for circum-neutral waters (Gasse *et al.*, 1995; Siver, 1999; Ramstack *et al.*, 2003). In terms of nutrient preferences, Siver (1999) derived an identical TN optima for both species (255 µg/L), as did Ramstack *et al.* (2003) for TP; 0.12 mg/L. In contrast, Tibby (2004) found differing TP optima for the two species (26.5 log<sub>10</sub> TP µg/L for *D. pseudostelligera* and 16.17 log<sub>10</sub> TP µg/L for *D. stelligera*), suggesting that fluctuations in nutrient supply may play a role in explaining the observed species shift in Lake Elingamite.

Thermal stratification also appears to be important to the ecology of both species so that they can maintain their position in the photic zone (Battarbee *et al.*, 2002; Rühland *et al.*, 2003; Karst-Riddoch *et al.*, 2005). In a study of a maar lake in south-eastern China, Wang *et al.* (2008) found that the abundance of *D. stelligera* decreased as winds increased and the lake became isothermal, while from Lake Redó in the Pyrenees, Catalan *et al.* (2002) found that the abundance of *D. pseudostelligera* increased when warmer temperatures were present during its growing months in autumn, at the end of the stratification process. Although no studies have been conducted into Lake Elingamite's stratification regime, it is unlikely that the lake stratifies in its current state due to its depth (3.4 m in 2003) and its exposure to wind. However, given the potential maximum depth of the lake (approximately 14 m), and the dominance of these stratification-dependent species in the pre-European period, it is highly likely that the lake stratified in the past.

Given that little separate these species in terms of stratification, nutrient, salinity or pH preferences, long-term changes in other solutes, such as silica, may provide a potential explanation for the gradual shift from *D. stelligera* to *D. pseudostelligera*. Smaller diatom species generally require less silica and, as discussed above, one of the morphological features that was used to separate these species was valve size, with *D. pseudostelligera* being the smaller of the two (*sensu* Krammer and Lange-Bertalot, 1991). Over millennial time-scales, some lakes have been found to become depleted in silica. After identifying a correlation between dissolved reactive silica and the width of sub-fossil sponge spicules, Kratz *et al.* (1991) reconstructed lake silica levels from eight lakes in Wisconsin, USA, over the Holocene. The authors found that, in all the study lakes, lake water silica content decreased over the period of the record. After comparing

against palynological evidence, they excluded climate as a cause of this decline. Instead, the authors propose that a loss of easily weathered material in the catchment, leading to a reduced silica transport to the lakes via groundwater, was the primary cause of the trends identified. There is insufficient evidence to support such a contention in the case of Lake Elingamite. However, in light of the discussion above and the highly similar resource and habitat requirements of the two species, the possibility of long-term silica decline can not be ruled out all together.

There are, therefore, only a few clues that can aid in the interpretation of the *D. stelligera* and *D. pseudostelligera* record from Lake Elingamite. It is possible that the gradually increasing dominance of *D. pseudostelligera* may indicate warmer autumns, but without similar knowledge on the life cycle of *D. stelligera*, this is only speculation. Similarly, the decrease of *D. stelligera* could be interpreted as indicative of increasing wind strength, though again, it is difficult to apply seasonally resolved ecological requirements to a record with sub-decadal sample resolution. According to the findings of Tibby (2004), different nutrient requirements may account for the shift in species, though this is not supported by the findings of other studies that find identical nutrient requirements for these species. Finally, it is also possible that long term lake-water silica decline may provide a potential explanation for the shifts in the relative abundances of these two species, though this is unable to be determined at this stage. It is clear though, that both species prefer a deep (> 8 m), clear, thermally stratifying lake, and while it can therefore be assumed that Lake Elingamite would have stratified in the past, the subtleties of the climate, or other, changes that are driving the shifts and trends between these species remain unresolved.

## **6.5 The regional climate of western Victoria since 500 AD**

The format of the following discussion is based on the ‘climatic zones’ identified in Figure 6.13 above. Where sample depths are noted for Lake Surprise, these relate to the corrected depth after correlation and merging of cores LSFS and LST1. Where ages are noted, the convention of using the term “AD” will be followed, this being equivalent to CE which denotes the Common Era, Christian Era or Current Era.



### 6.5.1 CZ4 – ca. 550 to ca. 680 AD

This period is characterized in the Lake Elingamite record by constant low conductivity values, suggesting that the lake was at, or near to, capacity throughout this 130 year period. As with much of the Lake Elingamite record, the diatom species assemblage over this period is dominated by the planktonic species *Discostella stelligera* and *D. pseudostelligera* which together, account for a maximum of 90.6% of the species assemblage around 615 AD, indicating a deep, clear, stratifying lake. The main feature of the Lake Surprise record is a peak in the inferred conductivity from 560  $\mu\text{S}/\text{cm}$  to 4437  $\mu\text{S}/\text{cm}$ , culminating at 650 AD. While this may be suggestive of a negative moisture balance, the diatom record during this period is dominated by *Cyclostephanos dubius* and *Stephanodiscus hantzschii*, neither of which are present in the modern training set. This accounts for an average squared-chord distance dissimilarity coefficient over this period of 165, which is greater than the value of the 20<sup>th</sup> percentile in the training set; 163.6. Therefore, this peak in inferred conductivity should be viewed with some caution.

The dominance of the planktonic *C. dubius* and *S. hantzschii* suggests a deep-water environment and, due to their nutrient preferences outlined above, a mixing regime recirculating nutrients. After 650 AD, sedimentary carbonate levels in Lake Surprise started to increase, indicating an increased importance of groundwater in the lake's hydrological budget. *Nitzschia paleacea* and *Synedra acus* dominate the species assemblage at this time, indicating low phosphorus levels relative to silica (Kilham *et al.*, 1986). This implies that less, or less vigorous, mixing of the lake was occurring, while blooms of non-silica based algae may have lead to the precipitation of carbonates. Prior to this, at around 565 AD in Lake Elingamite, *Fragilaria crotonensis*, a planktonic species indicative of high Si:P conditions (von Donk and Kilham, 1990) and nitrogen enrichment (Saros *et al.*, 2005), attains its maximum relative abundance (8.3%) in the record before slowly decreasing until it is completely absent.

The gradual demise of *F. crotonensis* culminates approximately 40 years before the increased conductivity that marks the upper boundary of the zone, and may indicate a subtle change in the mixing regime of the lake with less recirculation of nutrients. Following its demise, there is a decrease in the relative abundance of *D. stelligera*, and a corresponding increase in *D. pseudostelligera*, possibly suggesting that longer summers or warmer autumns (Catalan *et al.*, 2002) may have prevailed. Coincident with

these changes, *Fragilaria zeilleri* makes its first appearance in the Lake Elingamite record and continues its presence into the following zone. The presence of this non-planktonic species may indicate a gradual lowering of lake levels (*sensu* Tapia *et al.*, 2003; Ekdahl, *et al.*, 2008), though, if this is the case, it is not reflected in the modeled conductivity record. It is apparent though, that some gradual and subtle changes were occurring in the lake in the 50 years leading up to the rapid increase in conductivity in the following zone

In summary, the period of 550 – 650 AD was generally wet, characterised by full-lake conditions in Lake Elingamite. Both lakes stratified with nutrient recirculation evident in Lake Surprise. After 650 AD, groundwater became increasingly important to the hydrological budget of Lake Surprise, possibly due to changing P:E ratios caused by longer summers or warmer autumns. During this time, lake mixing in both lakes was decreased in either duration or strength. While DI conductivity in Lake Elingamite remained unchanged during this period, subtle changes in the diatom record are suggestive of a gradual change in climate after 650 AD.

#### 6.5.2 CZ3 – ca. 680 to ca. 900 AD

The changes evident during this climate zone are a notable feature of both lake records. Here, both lakes record their maximum inferred conductivity (excepting the post-European period in the Lake Elingamite record), indicating that a significant regional-scale climatic change took place.

Conductivity in both lakes increased after 680 AD. In Lake Surprise, this change was abrupt, highlighting the greater climatic sensitivity of the lake. In Lake Elingamite, a more gradual, though continual, conductivity increase occurred, culminating in a peak of 3923  $\mu\text{S}/\text{cm}$  at 776 AD. The gradual increase in conductivity is mirrored by the gradual increase in the proportion of benthic species in the assemblage, indicating that lake level was falling during this period (*sensu* Gasse *et al.*, 1989; Barker *et al.*, 1994a; Brugam *et al.*, 1998). This decrease can only be due to either less inflows and/or greater losses through evaporation or aquifer outflows (Mann, 1991). Given that this episode is also reflected in the Lake Surprise record, increased aquifer outflow can be discounted. Therefore, a major decrease in the P:E ratio has occurred, which persisted for almost a

century (680 AD –776 AD). Furthermore, the lake did not return to the baseline lake-full conditions until 867 AD, a further 91 years after peak conductivity was reached.

The increase in DI conductivity in Lake Elingamite is primarily driven by increases in the relative abundance of *Staurosirella pinnata*, *Fragilaria zeilleri* and *Psammothidium sacculum*, with coincident decreases in the relative abundances of *D. stelligera* and *D. pseudostelligera*. The reduction in the presence of these planktonic species is most likely due to a combination of decreasing lake depth and increasing conductivity, though the fact that they maintain some presence in the record indicates that conditions were still within the tolerance limits of these species. There are few discussions in the literature regarding the ecological preferences of either *F. zeilleri* or *P. sacculum*, so interpretation of their data is difficult. *Psammothidium sacculum* has been found in modern diatom assemblages from a 109 lake training set from northwest Canada and Alaska (Racca *et al.*, 2004), south-east Australian reservoirs (Tibby, 2004) and New Zealand lakes (Reid, 2005) though, in each instance, it was not found in sufficient abundance to determine tolerances to the differing environmental variables in the resulting transfer functions.

The ecology of *F. zeilleri* is similarly unclear. It was present in only three sites in the modern dataset and has a derived conductivity optimum of 3501  $\mu\text{S}/\text{cm}$  and a tolerance range from 1949  $\mu\text{S}/\text{cm}$  to 6287  $\mu\text{S}/\text{cm}$ . This optimum is markedly lower than the 3.03  $\log_{10}\text{S}/\text{cm}$  ( $\sim 10,071 \mu\text{S}/\text{cm}$ ) derived by Gasse *et al.* (1995), who also derived a pH optimum of 6.6. It has been alternately described as an “episammic taxon” (Tapia *et al.*, 2003:146), a “benthic diatom indicative of elevated salinity” (Ekdahl *et al.*, 2008:870) and a “freshwater epiphyte” (Liutkus and Ashley, 2003:697). Krammer and Lange-Bertalot (1991) offer no insight into this species’ ecological preferences, instead cautioning that, under a light microscope, this taxon can be difficult to distinguish from *Delphineis karstenii*, a marine species found in coastal regions of the Southern Hemisphere. While it does co-occur with *Staurosirella pinnata* at times in both fossil records, the patterns of co-occurrence indicate that they have differing ecological preferences. For instance, while these two species co-occur during the period of elevated conductivity in zone CZ3, they do not co-occur during the post-European period, when *S. pinnata* dominates. A similar pattern is evident in the Lake Surprise record as *S. pinnata* dominates during times of disturbance in zone CZ2 while *F. zeilleri* shows a muted response. While it is speculative, this may suggest that *F. zeilleri*

benefits during times when the conductivity is only moderate, yet the lake environment remains relatively undisturbed. However, with such a poor understanding of the ecology of both this species and *P. sacculum*, their value as indicator species is very low. Other species, such as *S. pinnata* for example, may provide a greater insight into the lake environment at the time.

*Staurosirella pinnata*, often referred to as one of the ‘Fragilarioides’ (e.g., Finkelstein and Gajewski, 2008), is a very common taxon. Considered to be an opportunistic species that prefers low-light, turbid conditions (Bouchard *et al.*, 2004; Antoniadou *et al.*, 2005; Karst-Riddoch *et al.*, 2005), it has very broad ecological tolerances (Van Dam *et al.*, 1994; Fluin *et al.*, 2009) making it a poor indicator species (Sonneman *et al.*, 2000). It has been described both as a benthic (Bennion *et al.*, 2001) and facultatively planktonic (Sonneman *et al.*, 2000) species. In Australia, *S. pinnata* often dominates wetlands with intermittent connectivity to rivers during times of high turbidity (Gell *et al.*, 2002, 2005; Leahy *et al.*, 2005). In the modern dataset it also appears to favour turbid conditions, dominating the surface sediment species assemblage from Lake Bolac for example (sampling depth, 1.4 m; secchi depth, 20 cm; conductivity 10,710  $\mu\text{S}/\text{cm}$ ), where it accounted for 68.1% of valves counted. This species maintains a constant presence in the fossil record from Lake Elingamite and is rarely absent from that of Lake Surprise. With such broad ecological tolerances, it is difficult to accurately interpret the root cause of fluctuations in its abundance and, therefore, *S. pinnata* can only be most accurately interpreted as being indicative of elevated disturbance and turbidity.

This interpretation is supported by the inferred environment in Lake Elingamite during the first century of this climatic zone in which lake levels are falling, benthos is becoming more dominant as a diatom habitat and, due to the lower water levels, turbidity is likely to be increased as lake mixing entrains bottom sediment. A similar pattern is also evident in the Lake Surprise record. Here, following the rapid demise of *Synedra acus* and *Nitzschia paleacea*, *Staurosirella pinnata*, *Staurosira elliptica* and *Fragilaria zeilleri* become more prevalent. It is also interesting to note an anomalous peak in the relative abundance of *Aulacoseira granulata*. This species is chain-forming and therefore requires turbulent environments to maintain its position in the photic zone (Kilham *et al.*, 1986). It also has high phosphorus and high silica requirements (Kilham

and Kilham, 1975; Kilham *et al.*, 1986). In the case of Lake Surprise therefore, the dominance of this species is indicative of a well-mixed environment.

The Lake Surprise DI conductivity record in this zone can be roughly divided in half. The earliest period has good modern analogues while the latter has very poor modern analogues. The division between these two periods aligns with the peak in conductivity in Lake Elingamite at 776 AD. As would be expected in lakes of differing sensitivities, the two records of increasing conductivity differ significantly. Lake Elingamite shows a very gradual increase, with a rise from 850  $\mu\text{S}/\text{cm}$  to 1252  $\mu\text{S}/\text{cm}$  occurring over 53 years, before a sudden increase from 1252  $\mu\text{S}/\text{cm}$  to 3923  $\mu\text{S}/\text{cm}$  in only 43 years, with a concurrent increase in the proportion of benthic species in the record from 45% to 87%. This sudden increase in conductivity suggests that either a threshold of lake sensitivity has been passed, whereby the smaller lake is more prone to evaporation than the larger lake, or that a natural threshold of species conductivity tolerances has been passed, leading to a sudden increase in taxa more tolerant of higher conductivities. A third interpretation may be a combination of these two factors.

Lake Surprise also demonstrates a step-wise increase in conductivity, albeit over a shorter time frame and with greater detail apparent. The initial conductivity increase in Lake Surprise occurs over a period of only 15 years, rising from 890  $\mu\text{S}/\text{cm}$  to 7715  $\mu\text{S}/\text{cm}$  between 680 and 695 AD. There followed a short-lived freshening episode, lasting only around five years, indicated by an anomalous peak in the relative abundance of *Aulacoseira granulata*, which accounts for 37% of the valves identified. As discussed above, this species requires well mixed environments and has high P and Si requirements (Kilham and Kilham, 1975; Kilham *et al.*, 1986). In the midst of a highly arid period, the most logical cause of turbulent, freshening conditions in conjunction with an increased supply of P and Si, is a brief, perhaps intense, humid phase delivering a greater sediment load into the lake. Assuming this interpretation is correct, the period of elevated rainfall was short-lived, with conductivity increasing again after this period, and, although it reached a peak of 9896  $\mu\text{S}/\text{cm}$  at 712 AD, both the magnitude and rate of change is less than the initial drought phase.

During this period of severe drought between 680 and 776 AD, carbonate laminations are present in the Lake Surprise record, indicating a strongly stratifying lake with an hypoxic hypolimnion. Following the peak in *A. granulata*, a very brief peak in

*Cyclostephanos dubius* and *Stephanodiscus hantzschii* is evident. Following the interpretive framework outlined in Section 6.4.1, these species bloom following a mixing event that recirculates nutrients in the lake. Yet here, they have bloomed during a time of carbonate laminations, which require a strongly stratified non-mixing lake for preservation. The cause of this short peak in the abundance of these species is therefore perplexing, though there are two possible explanations. The first is that this represents a single seasonal mixing event. This can not be ruled out since each one centimetre sample equates to approximately five years of sedimentation, and so seasonal resolution can not be determined. The second explanation is that aeolian or inwashed soil deposition has introduced nutrients to the system. Again, this is a plausible explanation as the peak in these species' abundance follows a period of severe aridity lasting *ca.* 30 years in which it could be expected that the extent of groundcover vegetation declined, falling lake-levels exposed the littoral zone, aeolian dust transport increased or a combination of these. Following this, the DI conductivity record indicates a short freshening episode, which corresponds with a decrease in the moisture content of the sediment. This suggests the transport and deposition of sediment with larger particle size and supports the hypothesis of an inwash of sediments and, therefore, nutrients. While both hypotheses provide plausible explanations for the results though, the underlying cause of this short-lived event can not be definitively resolved.

What is clear though is that the progression of species turnover, which up until this point largely followed the expectations of the resource relationship hypothesis of Kilham *et al.* (1986), seems to break down following the inception of the drought. During this time, the dominant taxa become the small fragilarioid group of *Staurosirella pinnata*, *Staurosira elliptica* and *Fragilaria zeilleri*. This may indicate that a degree of lake disturbance, as would be expected during such an arid phase, is occurring in the lake and it is now habitat preference, rather than nutrient availability, that is driving species turnover.

After 776 AD in Lake Surprise, and coincident with the decreasing DI conductivity in Lake Elingamite, the dominance of fragilarioid species decreases and there is a return of *C. dubius* and *S. hantzschii*. Though the DI conductivity at this time remains generally elevated, modern analogues are poor and confidence in this reconstruction is low. Indeed, the dominance of planktonic species, in conjunction with the freshening that is evident in Lake Elingamite, suggests that a less arid climate regime was in place than

that implied by the DI conductivity. However, there is evidence that a degree of climatic instability, which is not reflected in the Lake Elingamite record, remained.

Between 776 AD and 900 AD, the record of sedimentary carbonate content shows a pattern of regular fluctuations, which appear to occur on a quasi-bidecadal scale. This indicates that near-regular fluctuations between a high P/E climate and a low P/E climate were occurring. The regular rainfall events were sufficient to refill Lake Elingamite, which had returned to lake-full conditions around 890 AD. These rainfall events may also have been responsible for increased mixing in Lake Surprise, as the reappearance of *C. dubius* and *S. hantzschii* indicates an increased availability of nutrients, though this may also have been driven by inwash events after periods of low P/E. Between the rainfall events, the peaks in *N. paleacea* suggest that periods of decreased nutrient availability, possibly due to a depressed mixing regime (*sensu* Tracey *et al.*, 1996), were still occurring. These periods lasted up to a decade. It is also evident that during this time of instability, species turnover once again reverted to that of the resource requirements model (Kilham *et al.*, 1986), with peaks in the high Si:P *N. paleacea* following the dominance of the low Si:P *C. dubius* and *S. hantzschii*. This suggests that the relative importance of conductivity as a driver of species abundances may have been diminished, or that the tolerance of these taxa may have been expanded during these periods (*sensu* Saros and Fritz, 2000).

In summary then, this *ca.* 220 year climate zone can be divided into two sub-zones. During the first, from *ca.* 680 AD to *ca.* 776 AD, both lakes record a severe and prolonged drought which was possibly close to a century in duration. The onset of the drought was sudden, with conductivity in Lake Surprise increasing more than fourteen-fold in the space of *ca.* 30 years. Water levels at both lakes decreased and the diatom assemblages in both sites became dominated by benthic, opportunistic species. The Lake Surprise record suggests that there may have been a brief respite from the drought which appears to have caused increased runoff and deposition of topsoil and nutrients. During the second sub-zone, from *ca.* 776 AD to *ca.* 900 AD, planktonic diatoms returned to dominate the species assemblages of both lakes. Conductivity decreased in Lake Elingamite, and probably did so in Lake Surprise, yet a degree of climatic instability remained. The record of sedimentary carbonate content from Lake Surprise suggests that quasi-regular fluctuations between high P/E periods and low P/E periods occurred. The periods of high P/E were of a sufficient magnitude to gradually refill

Lake Elingamite and reinitiate the mixing regime in Lake Surprise. The periods of low P/E correspond with peaks in the high Si:P species (Kilham *et al.*, 1986) *Nitzschia paleacea*, indicating reduced mixing and nutrient circulation (Tracey *et al.*, 1996) occurred during these times.

### 6.5.3 CZ2 ca. 900 to ca. 1500 AD

The climatic instability that was evident in the upper part of the previous zone continues into this zone, however, both the amplitude and frequency of variation differs. In Lake Surprise, the beginning of the zone is marked by a dramatic decrease in the number of planktonic species in the diatom record, which falls from 86.5% to 0% between 900 AD and 944 AD. Initially, as was the case before, *N. paleacea* dominated after the decline of *C. dubius* and *S. hantzschii* at the end of the previous zone. However, as numbers of *N. paleacea* began to decline soon after 900 AD, *Staurosirella pinnata* became more dominant. This species shift coincides with increasing sedimentary carbonate content and increasing DI conductivity, indicative of a drier climate, while the dominance of *S. pinnata* indicates a disturbed, probably turbid environment (e.g., Gell *et al.*, 2002, 2005; Leahy *et al.*, 2005).

Carbonate laminations are evident in the sediment and show a sharply delineated lower boundary (Figure 5.19), signifying a sudden onset of hypolimnetic anoxia and groundwater dominance. Laminations are present in the sediment for the next 11 centimetres of the core, during which time *S. pinnata* steadily increases its dominance until it accounts for 52% of the species assemblage at 971 AD. During this period, *Encyonopsis ruttneri*, *E. microcephala*, *Gomphonema affine* and *Achnantheidium minutissimum* also increase their prevalence, suggesting an increase in lake water pH (Kelly, 2000; Sonneman *et al.*, 2000). The continual presence of laminations between 922 AD and 981 AD, and the absence of low Si:P species at this time, indicates that lake mixing was severely limited. This, in turn, implies that the lake probably stratified for longer periods of the year.

Surprisingly, there is scant evidence of this increasing aridity in the Lake Elingamite record. Inferred conductivity remains low, though it does gradually increase from 857  $\mu\text{S}/\text{cm}$  to 1170  $\mu\text{S}/\text{cm}$  between ca. 950 AD and 1005 AD, indicating a very slight decrease in lake levels. Both *Discostella pseudostelligera* and *D. stelligera* remain



dominant, though the relative abundance of *S. pinnata* gradually increases from 6.5% to 18.5% during this period. Thereafter, the abundance of this species decreases until it accounts for only 7% of the assemblage at *ca.* 1040 AD. This change is roughly coincident with a rapid decrease in DI conductivity in Lake Surprise in which carbonate laminations cease and sedimentary carbonate content falls from an average of 23% to 7%. Concurrent with this is an increase in both the moisture and organic content of the sediment, signifying a deep lake environment with fine sediment particles of probable authigenic origin.

Interestingly, despite a return to deep-lake conditions, it is *Discostella pseudostelligera* and, to a lesser extent, *D. stelligera* that appear in the record, rather than the *C. dubius*/*S. hantzschii* combination that was so prevalent before. This represents the first appearance of the two *Discostella* species in the record, and, while it is a short-lived appearance in only two samples (equating to approximately a decade), *D. pseudostelligera* dominates the species assemblage at this time, accounting for 38% and 40% of all valves counted. Why it was this species and not *C. dubius* that benefitted at this time is not clear, though it could be speculated that, given their differing trophic preferences, little nutrient recirculation was occurring.

After this short period dominating the Lake Surprise species assemblage, the relative abundance of *D. pseudostelligera* falls away very rapidly as sedimentary carbonate content and conductivity increase. Around *ca.* 1052 AD, *S. pinnata* attains its maximum relative abundance in the record, accounting for 82.5% of all valves enumerated. From this point onwards, until *ca.* 1250 AD, there are numerous and rapid fluctuations in all measured and inferred variables in the Lake Surprise record. Diatom-inferred conductivity shows several peaks and troughs, corresponding to the often rapid fluctuations in diatom species abundances. There are a number of periods, of varying time-spans, where carbonate laminations were deposited. As these periods cease, there is an immediate reappearance of *C. dubius* and *Stephanodiscus hantzschii*, signifying the recommencement of a vigorous mixing regime in the lake and the recirculation of nutrients.

The rapid turnover of species indicates the changing lake environment during this time. The climate during this period appears to have been highly variable with alternating arid and wet phases lasting between 15 and 30 years. The relative abundance of *S. pinnata*

peaks approximately every 20 – 25 years and is the primary cause of the increases in DI conductivity. During the intervening wet phases, *Synedra acus* and/or *Nitzschia paleacea* return to dominance. The fluctuations in DI conductivity and sedimentary organic, carbonate and moisture content, indicate rapid oscillations in lake levels. However, species in the low Si:P category (Kilham *et al.*, 1986), such as *C. dubius* and *S. hantzschii* mentioned above, do not regularly reappear during the fresh, deeper-lake phases. This suggests that, while the lake was probably holomictic, especially during times where carbonate laminations are absent, vigorous remixing rarely occurred.

Also of interest is the first significant appearance of *Cyclotella meneghiniana* in the Lake Surprise record around 1100 AD. This planktonic species is tolerant of high nutrients (Kelly and Whitton, 1995) and is considered to be a facultative nitrogen heterotroph (Van Dam *et al.*, 1994). It has also been found to be tolerant of a wide conductivity range (e.g., Gell, 1997; Davies *et al.*, 2002), a finding reflected in this study with a derived tolerance range of 1802  $\mu\text{S}/\text{cm}$  to 10,117  $\mu\text{S}/\text{cm}$ . As such, its value as a conductivity indicator is poor. Its occurrence coincides with a period of decreased sedimentary carbonate content and increased percentage of organic and moisture content, thus implying a humid climate and a deep lake environment. There are three main peaks in the relative abundance of this species between 1200 AD and 1250 AD and, in each instance, the peak also coincides with a decrease in carbonates and an increase in organic and moisture content. The Lake Elingamite record during this time indicates decreasing conductivity, lending support to the interpretation that this period was humid.

It is unclear what was driving the appearances of *C. meneghiniana* in Lake Surprise. Its co-occurrence with seemingly wet periods and its ability to thrive by utilising nitrogen could suggest either greater delivery of atmospheric nitrogen through increased precipitation, or greater catchment runoff delivering more nitrogen to the lake in the form of particulate organic nitrogen (Cohen, 2003). In the case of the latter though, this seems unlikely given the small catchment area. Another scenario could be that nitrogen is being developed as a by-product of primary production. Saros and Fritz (2000:451) note that through nitrogen fixation, the presence of blue-green algae can raise N:P ratios by converting “N from  $\text{N}_2$  into forms usable by other algae” and in doing so, can contribute “anywhere from 0-82% of the total open water N budget.” Without

supportive data though, it can not be determined if this was the case in Lake Surprise and the underlying driver of this taxon's abundance remains unclear.

The general interpretation from Lake Elingamite matches that of Lake Surprise over this time-span, with fluctuations in DI conductivity evident. In line with the less sensitive nature of this lake, the amplitude of variations is more muted than in Lake Surprise. Even so, the correspondence between the two records is strong. Despite the obvious presence of some prolonged arid periods, there is no return to the elevated conductivity evident in the previous zone. This suggests that the intervening wet periods, though often short in duration, may have been of a significant magnitude. The DI conductivity does, at times (e.g., *ca.* 1150 – 1250 AD), double, signifying a considerable alteration to the lake environment, though generally, conductivity in Lake Elingamite remains low. From *ca.* 1150 AD, there is a gradual change evident in the diatom flora of Lake Elingamite. There is a steady decline in the relative abundance of *Discostella stelligera* and a coincident rise in *D. pseudostelligera*. Shortly after, *Staurosirella pinnata* peaks before its abundance also drops away. There is a steady increase in the relative abundance of *Staurosira construens* var. *construens* which peaks from 1350 – 1400 AD. This variety of *S. construens* was not recorded in the modern dataset, though constrained ordination of species and environmental variables from Murray River wetlands suggest that it prefers slightly fresher, and possibly slightly more turbid, environments than *Staurosirella pinnata* (Gell *et al.*, 2002). The steady increase in the relative abundance of this species may therefore be indicative of a gradual shift towards a wetter climate, with increased transport of sediment from exposed crater walls. By extension, its sudden demise around 1400 AD, in conjunction with the increased abundance of the clear- and deep-water planktonic species *D. stelligera* and *D. pseudostelligera*, could indicate that full-lake conditions were attained with less exposed crater walls delivering less sediment. This is a speculative interpretation, though the rapid increase in abundance of *Staurosira construens* var. *construens* in the post-European period provides tentative support.

The upper boundary of this zone is located at roughly 1500 AD and, while the two DI conductivity records do not initially seem to record this boundary synchronously, much of the Lake Surprise record between 1250 AD and 1500 AD has poor modern analogues, making the reconstructions less reliable. However, coincident with the zonal boundary implied by the Lake Elingamite record is a clear shift in Lake Surprise to a

greater representation of planktonic species in the record, with *Cyclostephanos dubius* in particular rapidly increasing in relative abundance (Figure 6.6). This indicates that a shift in climate at this time is recorded in both lakes.

In summary, the diatom records from both lakes in this climate zone are representative of a highly unstable climate. The greater detail that is evident in the Lake Surprise record indicates numerous, rapid increases in DI conductivity, interspersed with equally rapid returns to low conductivity values. Similarly, the sediment analysis shows numerous, rapid fluctuations in all measured variables. In particular, the period from 900 – 1250 AD, though possibly extending to *ca.* 1400 AD, seems exceptionally variable with shifts between well-mixed, deep-lake environments with high sedimentary organic and moisture content, to shallow lake environments with hypolimnetic anoxia leading to the preservation of laminations at times. The rapid shifts between positive and negative moisture balance, at sub-decadal to decadal frequency, is suggestive of fluctuations between *El Niño* and *La Niña*-like conditions, or of the varying influence of the Indian Ocean Dipole. These possibilities are explored further in Section 6.7 below.

#### 6.5.4 CZ1 *ca.* 1500 to *ca.* 1875 AD

As mentioned above, the boundary of this zone is marked by an increased prevalence of planktonic species in Lake Surprise and the cessation of climatic variability that typified the previous zone. In Lake Elingamite, the zone is characterised by persistently low inferred conductivity values averaging only 800  $\mu\text{S}/\text{cm}$  (minimum 696  $\mu\text{S}/\text{cm}$ , maximum 1029  $\mu\text{S}/\text{cm}$ , standard deviation 74.7  $\mu\text{S}/\text{cm}$ ) over this 375 year period. When compared to the gradual freshening that is evident in Lake Surprise, this suggests that the lake was either full, or near to full, during this time. In Lake Surprise, sedimentary variables are all relatively stable for the majority of this zone, with generally high organic and moisture content and correspondingly low carbonate content. Yet, over the background of this long-term freshening of Lake Surprise, there is evidence of climatic variability within the diatom record.

In Lake Surprise, CZ1 commences with two distinct peaks in the relative abundance of *Cyclostephanos dubius* centered around 1510 and 1565 AD, with an increase in the representation of *Cyclotella meneghiniana* in the intervening period and directly after. It is interesting that the peak in the low Si:P species *C. dubius* is not followed by an

increase in the relative abundance of a high Si:P species (*sensu* Kilham *et al.*, 1986) as it was in earlier portions of the record. Instead, for a period of approximately 120 years, there are fluctuations between the two planktonic species *C. dubius* and *Cyclotella meneghiniana*.

The mechanisms driving these shifts in species can only be speculated upon and little information can be garnered from the conductivity reconstruction as the modern analogues are poor for much of this period. The interpretive model for Lake Surprise (Section 6.4.1) proposes that the appearance of *Cyclostephanos dubius* is dependant upon the presence of elevated phosphorus levels which, in turn, indicates lake mixing and nutrient recirculation. The causes of increasing relative abundance of *Cyclotella meneghiniana* though, remain unclear. However, given that *Cyclotella meneghiniana* is a nitrogen heterotroph (Van Dam *et al.*, 1994), fluctuations in the N:P ratio may be the most likely cause of the observed variation in relative abundances of these species within this *ca.* 120 year period (~1510 – 1630 AD). The mechanism, or mechanisms, driving the proposed changes in N:P ratios can not be determined and the broad conductivity tolerance of *C. meneghiniana*, in conjunction with the absence of *Cyclostephanos dubius* from the training set, does not help clarify the situation. There is also little insight into any climatic fluctuations of the time provided by the Lake Elingamite record, where persistently low DI conductivity values indicate a full to overflowing lake.

Following the period of gradually increasing relative abundance of *Cyclotella meneghiniana* in the Lake Surprise record, numbers rapidly fall away and *D. pseudostelligera* and *D. stelligera* dominate the record, attaining their maximum relative abundances of 57.9% and 29.6% respectively, between *ca.* 1638 AD and 1648 AD. The combined dominance of these species drive the inferred conductivity down to 709  $\mu\text{S}/\text{cm}$  during this decade and indicate the presence of an oligotrophic lake of substantial depth. Both species maintain a decreased presence in the diatom assemblage for the rest of this zone, indicating that a stratification regime was present in the lake (Battarbee *et al.*, 2002; Rühland *et al.*, 2003; Karst-Riddoch *et al.*, 2005), with differing degrees of mixing evident in the fluctuating prevalence of *Cyclostephanos dubius*. There are small peaks in inferred conductivity around 1660, 1700, 1750 and 1820 AD, though the general trend indicates increasing lake depth until *ca.* 1880. This corresponds with reports that other lakes in the region, such as Lake Bullenmerri, were

full and/or overflowing in the mid-1800s (Currey, 1970). It is also consistent with the findings of Jones *et al.* (2001) who, using data from Lake Bullenmerri and other western Victorian lakes, state that the high lake levels at the time were “maintained by a climate that was almost certainly wetter, probably cloudier and perhaps cooler, when compared to the instrumental climate record” (Jones *et al.*, 2001:178).

Sedimentary analysis, in particular organic and carbonate content, indicates rapid fluctuations above a depth of approximately 110 cm, recorded in core LSFS. As discussed in Section 5.2.1.1, these fluctuations are likely to be an artifact of smaller sample weights and should be viewed with caution. Measurements of moisture content in the upper portion of the combined record also appear spurious when the records from the two cores are combined, being markedly higher than at any point in the remainder of the record. As mentioned previously, this is due to the frozen spade coring method employed for core LSFS which freezes sediment and water to the plates of the corer.

In summary, the climate during the time covered by this zone (*ca.* 1500 – 1875 AD) is generally characteristic of a period of positive moisture balance. Inferred conductivity in Lake Surprise shows a continual and gradual decline, while Lake Elingamite appears to have been full to overflowing for much of this time. During this period of increasing moisture balance, the Lake Surprise diatom record indicates periods of increased precipitation and differing degrees of lake mixing, though in comparison to the previous zones, there is marked stability in the climate regime of the time.

#### 6.5.5 The post-European period – *ca.* 1875 to present

As discussed above, the fossil diatom record of Lake Elingamite in the post-European period provides little in terms of climatic information. The diatom assemblage is dominated by *Staurosirella pinnata*, *Staurosira construens* var. *construens* and *Cocconeis placentula* and the diatom-inferred conductivity reconstruction shows no relationship with instrumental climate records over the 20th century (see Figure 6.9 above) or other accounts of lake depth. The modeled (Mann, 1991) and measured (Hussainy, 1969) lake levels for instance, as well as eye-witness accounts (Jim Bowler, pers. comm.), all indicate that the lake was near to overflowing up until the mid-1960s. However, the diatom record does not reflect this, nor the dramatic fall in water levels since that time. The fossil diatom assemblage during this time is therefore considered to

be driven by the impacts of vegetation clearance and landscape disturbance wrought by the arrival of Europeans, and any climate signal in the record is masked.

In contrast, the record from Lake Surprise provides detail which, as noted above, reflects changes in the P/E ratio as expressed by the PDSI of Ummenhofer *et al.* (2009). Soon after 1870, the relative abundance of *Discostella pseudostelligera* decreases, possibly indicating falling lake levels and/or a shift in the climate, as proposed by Jones *et al.* (2001). Thereafter, from approximately 1895 to 1915, the dominant taxon is *Cyclostephanos dubius*, indicating an increased mixing regime. Despite the dominance of this planktonic species, lake levels may have continued to decline during this time, as non-planktonic species, namely *Staurosirella pinnata* and *Staurosira elliptica*, begin to increase in prevalence after approximately 1905, possibly in response to the 'Federation drought' which occurred between 1895 and 1902 (Ummenhofer *et al.*, 2009). These species maintained their dominance until approximately 1940, after which *Cyclostephanos dubius* and *Cyclotella meneghiniana* began to increase in relative abundance. Allowing for the approximate 6 year age offset discussed in Section 6.2, this increase in abundance of planktonic forms is coincident with a shift towards positive rainfall anomalies (Figure 3.3).

Through the 1950s, the dominance of *Cyclostephanos dubius* and *Cyclotella meneghiniana* alternate, suggesting fluctuating N:P ratios. Both species decrease their relative abundances around 1960 and there is a coincident increase in the relative abundance of *Synedra acus*, illustrating a shift from low Si:P to high Si:P species (*sensu* Kilham *et al.* 1986) as was evident earlier in the record. From the mid-1960s though, and despite periods of positive rainfall anomaly (Figure 3.3), *Staurosirella pinnata* began to increase in dominance, a trend that continues to the surface of the core, in line with the trend of temperature anomalies of the region (Figure 3.4).

Though Lake Surprise continues to exhibit an annual stratification and mixing regime (Timms, 1975), it is interesting to note that species such as *C. dubius* and *Stephanodiscus hantzschii* are not evident in the contemporary setting. Instead, since the 1970s, the diatom flora has been dominated by small fragilarioid species; *Staurosirella pinnata*, *Staurosira elliptica*, *Pseudostaurosira brevistriata*, *Frankophila similioides*, *Fragilaria zeilleri* and *Fragilaria polonica*. This is consistent with the findings from both lakes during previous arid phases when fragilarioid species dominate the diatom

assemblages. However, the key difference between the lakes in the post-European phase is that the DI conductivity record in Lake Surprise has been shown to reflect moisture balance. With this established, the present-day drought in south-eastern Australia can be viewed in an historical context.

## **6.6 Drought frequency, intensity and duration**

Droughts detrimentally impact upon the environment, society and economy of afflicted areas and, in the southeast of Australia, some impacts of the current drought are unprecedented in historical records (Murphy and Timbal, 2008). Despite previous droughts in the region having a similar duration (e.g., 1936 – 1945 drought), the present drought has a greater intensity, which has been attributed to rising temperatures over recent decades (Nichols, 2004; Ummenhofer, 2009). In addition, while droughts are a natural characteristic of Australian climatic variability (Ummenhofer, 2009), they are predicted to become more common as global warming continues (Mpelasoka *et al.*, 2008).

The fine-resolution record of the climatically sensitive Lake Surprise allows the current drought to be placed into an historical perspective in terms of its severity and duration, while also informing the variability of drought frequency over the last *ca.* 1500 years. In order to illustrate the history of drought frequency, intensity and duration, the diatom-inferred conductivity record is presented in terms of deviation from the average of the entire conductivity record ( $n = 344$ ; Average = 3610  $\mu\text{S}/\text{cm}$ ; S.D. = 1816  $\mu\text{S}/\text{cm}$ ), an approach employed by Laird *et al.* (1996a, 1996b, 2003). The same approach was applied to the Lake Elingamite record, though in this instance, the average of the conductivity was calculated after excluding the post-European samples ( $n = 55$ ), as these are considered not to be reflecting climate forcing (Section 6.2). After exclusion of these samples, the average inferred conductivity of the remainder of the Lake Elingamite record ( $n = 240$ ) was 1029  $\mu\text{S}/\text{cm}$ , with a standard deviation of 372  $\mu\text{S}/\text{cm}$ . This record is provided for comparison, with the following discussion primarily focusing on the more sensitive Lake Surprise record.

Given that the Lake Surprise chronology is derived from a limited number of dates, there are obvious qualifications in terms of interpreting the precise duration of the droughts. However, the general timing of drought episodes is generally well supported



by corresponding changes, albeit in muted form, in the Lake Elingamite record. Portions of the Lake Surprise diatom record have poor analogues in the training set and the reconstructions from these parts of the record remain questionable. In order to present valid comparisons of, and between, droughts then, only samples with “good”, or “very good”, modern analogues are considered. It should also be noted that not every period of positive deviation will represent a period of drought, but may merely be a slight deviation away from the average condition. For the purposes of this discussion, only periods that exceed one standard deviation of the conductivity record are considered severe enough to be representative of major drought. It is, however, recognised that this is not a comprehensive definition of drought periods. For example, despite periods of twentieth century drought (e.g., Ummenhofer *et al.*, 2009) being evident in the Lake Surprise conductivity record, only the current drought deviates from the average by more than one standard deviation. Hence, by excluding periods with conductivity less than one standard deviation from the average of the record, only the most severe drought periods will be captured.

The drought reconstructions are presented in Figure 6.14 and clearly demonstrates that frequent droughts, of a greater severity than the current drought, have occurred in the past. Specifically, the period between *ca.* 680 AD and 770 AD is exceptionally dry with three distinct and very severe drought episodes, of possibly multi-decadal duration, evident in the Lake Surprise record. Due to the lower sensitivity of Lake Elingamite, only the cumulative effect of these periods is recorded, though this 90 year period represents the driest phase of the records which is of an apparently greater severity than the contemporary drought.

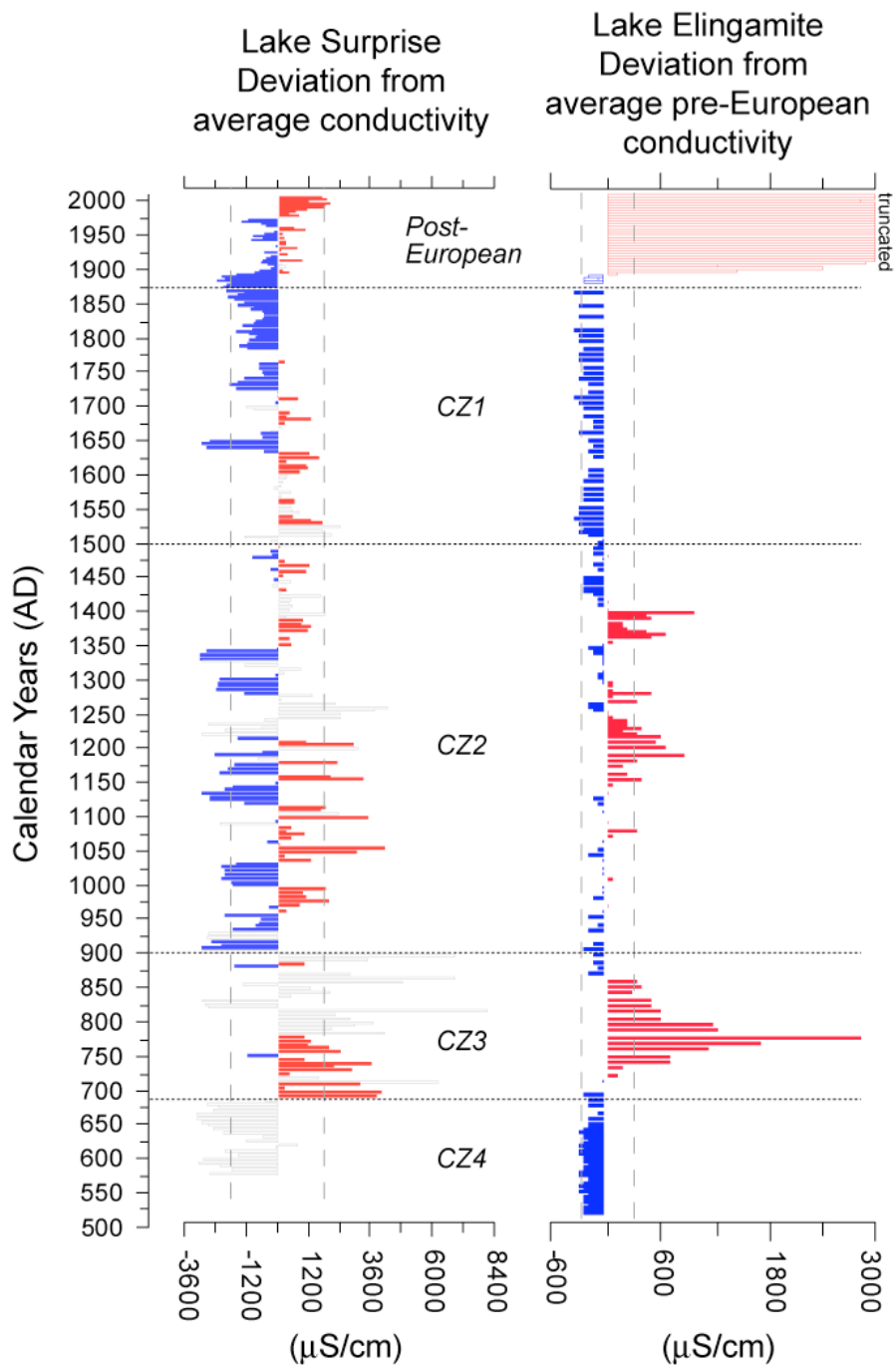


Figure 6.14: Deviation from the average inferred conductivity of the entire Lake Surprise record (left) and the pre-European portion of the Lake Elingamite (right) record. Grey bars in the Lake Surprise record indicates samples with poor modern analogue. Hollow bars in the Lake Elingamite record are samples that were excluded when calculating the average of the inferred conductivity (see text for details). Dashed lines equal  $\pm 1$  standard deviation.

The presence of droughts more severe than the current drought is seemingly at odds with the results of Gill (1971), and Bowler and Hamada (1971), who dated drowned trees emerging from falling lake-levels in Lake Bullenmerri and Lake Keilambete

respectively. Their results dated the trees at  $1865 \pm 85$  and  $1890 \pm 115$   $^{14}\text{C}$  years, which, after calibration<sup>6</sup>, indicates ages of 220 AD for Lake Bullenmerri and 170 AD for Lake Keilambete. As the trees would have deteriorated had they been exposed after drowning, this signifies that lake levels, in these two lakes at least, had been greater than 1971 levels for around 1800 years. This places them well outside the period of the drought identified in this study (approximately 680 – 770 AD) and may raise questions regarding the inferred severity of the drought which was identified in the records from both study sites.

As the trees emerged more or less synchronously from two different lakes that are approximately 20 km apart, the indications are that declining lake-levels are due to climate rather than site-specific factors. However, different lakes respond differently to differing climate forces (e.g., Magnuson *et al.*, 2004). For example, from an examination of historical records, Jones (1999) finds evidence of an apparently severe drought in western Victoria from approximately 1838 – 1843. During this time, Lake Bolac, a large ~1400 ha lake (Jones, 1999) was noted to have completely dried, an event, in the current drought, which only occurred in 2009 (Peter Gell, pers. comm.). However, despite the apparent severity of the drought in the early 1840s, it was not severe enough to expose the drowned trees in Lakes Keilambete and Bullenmerri. Another interesting point to note, is that at the time of the trees emerging from these lakes, the climate of the preceding decades was the wettest of the 20th Century (Figure 3.3) with low mean temperature anomalies (Figures 3.4), which are reflected by a strongly positive PDSI (Figure 6.10). The implication therefore, is that the lake-level decline in Lakes Bullenmerri and Keilambete, is a response to a longer-term fall in effective precipitation, and that these lakes are not as responsive to shorter-term fluctuations in moisture balance. Jones *et al.* (2001) notes that the depth of Lake Keilambete has been declining since the first measurements were taken in 1859 and, from a study which included also two other western Victorian lakes, concluded that a change in the climate of the region occurred some time between 1840 and 1863 which decreased the P/E ratio. This appears the most likely mechanism for the gradual, and continual, decline in lake levels which eventually exposed the drowned trees.

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<sup>6</sup>Dates were calibrated in the program Calib (version 5.0.1) using the SHCal04 calibration curve (McCormac *et al.*, 2004). Ages were determined by taking the median ages of the 2-sigma age ranges. Lake Bullenmerri; 2-sigma age range 25 – 414 AD, probability distribution = 1. Lake Keilambete; 2-sigma age range BC 95 – 414 AD, probability distribution = 0.992.

From *ca.* 900 to 1200 AD, the appearance of drought becomes less frequent, and while the droughts that do occur are of a notable severity, it appears that the duration of these droughts was not sufficient to expose the drowned trees in Lakes Bullenmerri and Keilambete for long enough to lead to their deterioration. In addition, the intervening humid periods appear to be of a longer duration than the majority of the drought periods, highlighting the variable nature of the climate at the time, as discussed above. The timing of this period of the record falls within the commonly identified boundaries of the Medieval Warm Period in Europe (discussed further in Section 6.8 below).

From *ca.* 1350 to 1625 AD, there are frequent positive deviations from the average conductivity, though none of these periods exceed one standard deviation and none are of an intensity similar to that of the contemporary drought. Indeed, prior to the onset of the current drought, no period of comparable drought intensity has occurred since *ca.* 1200 AD, highlighting the significance of the event. Also of note is the lengthy period prior to *ca.* 1875 AD which is dominated by negative deviations from the average. This corresponds with a period from *ca.* 1600 AD to the late 1800s in which Tasmanian tree-rings infer decreased temperatures during the spring-summer growing season (Cook *et al.*, 2000).

From the 1830s onwards, the study region was being intensively explored for suitable agricultural land (Jones, 1999). On the basis of the evidence presented here, it appears that this was occurring during unusually humid times, which may have given the original settlers a false impression of the nature of the landscape. The drought of the 1840s, mentioned above (Jones, 1999), is reflected in the Lake Surprise conductivity deviation record (Figure 6.14 above) as a slight positive deflection soon after ~1825, followed by increasingly negative values around 1850, by which time Lake Bolac had re-filled (Jones, 1999). There is no evidence of this period in the Lake Elingamite record, thus highlighting the comparative sensitivity of Lake Surprise. In the context of the extended period of positive moisture balance in which it occurred, the 1840s drought appears to be of little hydrological importance.

The Lake Surprise record indicates that, while the drought that is currently affecting the study area is clearly an unusual and uncommon event, it is not unprecedented. In terms of intensity, the current drought is unmatched for almost 800 years, and even then, the droughts were seemingly of a shorter duration. Indeed, the record indicates that the last

drought of a similar duration to the contemporary event occurred around 1200 years ago. It also appears that lake levels, in Lakes Keilambete and Bullenmerri at least, have not been at similar levels to present for well over a millennium, thus placing the current drought in stark contrast.

## **6.7 Comparisons and synthesis with other studies**

This study aimed to reconstruct climatic fluctuations in western Victoria in fine-resolution, thereby addressing a dearth of research on the topic. As has been demonstrated above, it has been possible to assemble a great deal of information by examining samples with sub-decadal temporal resolution, with individual drought events among the detail identified. However, comparing the results of this study to others from the same region is made difficult by the very lack of fine-resolution records that spawned this study. Indeed, the only fine-resolution (that is, each sample equating to less than a decade), short-term studies from nearby are those of Cook *et al.* (1991, 1992, 1995, 1996), the culmination of which is Cook *et al.* (2000), a tree-ring study of growing season temperatures from western Tasmania. While this is instructive, the climate drivers on western Tasmania differ from those of the southeast of mainland Australia. For example, the key driver of dry conditions in western Tasmania has been identified as the positive phase of the Southern Annular Mode (SAM) (Hill *et al.*, 2009) while in the southeast of the mainland, the Indian Ocean dipole and ENSO are key drivers of aridity (Ummenhofer *et al.*, 2009). However, as this is the only high-resolution study from nearby, and given that the SAM also influences the climate of the study region (see Section 3.2.4), the Cook *et al.* (2000) study is taken into consideration. In order to gain a broader understanding of the forces driving the climatic changes evident in the records though, there is a need to also examine studies from further afield.

In a detailed examination of the extent, coherence and causes of rapid climate change events during the Holocene, Mayewski *et al.* (2004) compiled the results of numerous climate studies that had highly resolved sampling resolution, high dating quality and were from a broad global distribution (though notably none were from Australia or New Zealand). The authors identified periods of rapid climate change based on the glacier fluctuation record of Denton and Karlén (1973) and used the Greenland Ice Sheet Project Two (GISP2) chemistry series (Mayewski *et al.*, 1997) to verify the chronology. These periods were utilised as a framework of the study as they represent periods which

contain “geographically broad evidence for change in Holocene climate” (Mayewski et al., 2004:244). Mayewski *et al.* (2004:246-247) assert that because of the “globally distributed signature” of these events, they can be considered to be “of worldwide significance.” Of specific interest to this project are the two such periods that occurred within the last two millennia at 1200-1000 and 600-150 cal. yr B.P. These are illustrated in the comparisons made below in order to compare the results of this study against those from a broader geographical context (Figure 6.15). Other southern hemisphere studies, of varying proxies, are identified for inclusion in the comparisons, based on their ability to inform the broader climatic influences of the study region or because of their geographical proximity to the study site. A number of informative studies are not included in the ensuing illustration, usually because the results do not lend themselves to graphic representation. These studies are referenced separately in the text.

The comparisons between the various climate records are illustrated in Figure 6.15 and it is apparent that there is a clear coherence between the DI conductivity record of Lake Surprise (where good modern analogues exist) and the reconstruction of *El Niño* frequency from Ecuador (Moy *et al.*, 2002). The earliest period of notable climatic change in the Lake Surprise record is a period of increased drought frequency, severity and duration centered around 725 AD. This is the driest period of the record and corresponds with the maximal number of *El Niño* events per century ( $n = 31$ ) in the last 2000 years (Moy *et al.*, 2002). Tasmanian temperatures at this time show an increased frequency and amplitude of variability (Cook *et al.*, 2000) and there is also a peak in Southern Ocean temperatures, possibly due to a decrease in sea-ice cover and a poleward migration of the westerlies (Rosqvist and Schuber, 2003). Mooney (1997) identifies a period of warmer temperatures around this time, as inferred from minerogenic and pollen changes from Lake Keilambete in western Victoria.

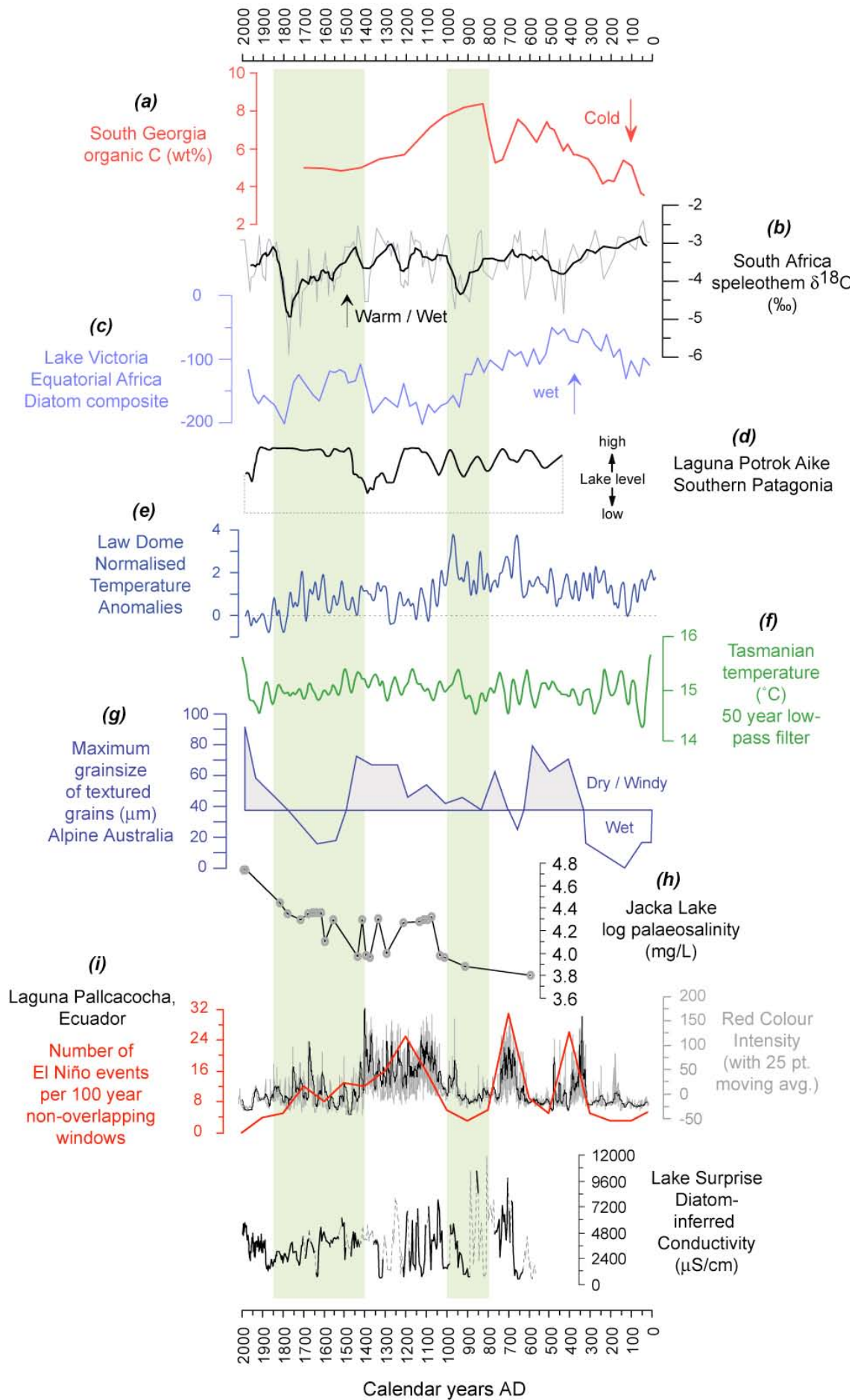


Figure 6.15: The Lake Surprise diatom-inferred conductivity record compared against other studies. Dashed sections of the Lake Surprise record indicate periods with poor analogues in the modern calibration set. Green bands represent periods of globally significant rapid climate change as defined by Mayewski *et al.* (2004). Figures (a)-(c) cited in and redrawn from Mayewski *et al.* (2004). (a) Sedimentary organic carbon percentage from Georgia, Southern Ocean (Rosqvist and Schuber, 2003). (b)  $\delta^{18}\text{O}$  record (‰) for speleothem in Cold Air Cave, South Africa (Lee-Thorp *et al.*, 2001). (c) Lake level proxy based on diatom ratio from Lake Victoria, Africa (Mayewski *et al.*, 2004). (d) Multi-proxy lake level reconstruction from southern Patagonia, Argentina (Haberzettl *et al.*, 2005). (e) Law Dome temperature anomalies, normalized over the 1751-1950 period and smoothed with a 50-year Gaussian filter. Redrawn from Jones and Mann (2004). (f) Tree-ring based warm season temperature reconstruction from western Tasmania (Cook *et al.*, 2000). (g) Aeolian grain size from Blue Lake, Snowy Mountains, Australia. Redrawn from Stanley and De Deckker, (2002). (h) Ostracod-inferred salinity reconstruction from Jacka Lake, western Victoria, Australia (Radke, 2000). (i) Red-scale intensity and El Niño events per century from Laguna Pallcacocha, southern Ecuador (Moy *et al.*, 2002).

The record of Stanley and DeDeckker (2002), which also demonstrates coherence with the Moy *et al.* (2002) record, indicates a shift towards drier and windier climates around 800 AD. The dust source for that study differs to the Lake Surprise study region (discussed further below), indicating that aridity at this time extended beyond western Victoria and had a broad spatial impact. The correspondence between the record of Stanley and DeDeckker (2002), Moy *et al.* (2002) and that from Lake Surprise, differs by approximately 50 years, possibly due to the low sampling resolution of the Stanley and DeDeckker study. The ostracod-inferred conductivity record from Jacka Lake in western Victoria (Radke, 2000) does not record this event, though again, this may be a function of a low sampling resolution of only two samples between 1000 and 1500 cal. yr B.P.

It appears therefore, that periods of heightened *El Niño* activity evident in lake deposits in Ecuador are also reflected in the palaeoclimate of western Victoria in the form of severe and frequent drought episodes. The reconstruction of *El Niño* frequency from Laguna Pallcacocha only records moderate to strong *El Niño* events (Moy *et al.*, 2002), and events of this magnitude have been shown to modulate the Southern Annular Mode (L'Heureux and Thompson, 2006), possibly impacting upon the climate recorded in the Southern Ocean at this time (Rosqvist and Schuber, 2003). Ummenhofer *et al.* (2009) found that droughts in south-eastern Australia were more severe when *El Niño* events occurred in conjunction with positive Indian Ocean Dipole (IOD) events (cool eastern pole, warm western pole). This may have been the situation during this period of aridity



and, although there are presently no millennial-length reconstructions of IOD against which to compare, an examination of all ENSO and IOD events between 1876 and 1990 found positive IOD events only co-occurred with *El Niño* and ‘no ENSO’ events and never with *La Niña* events (Meyers *et al.*, 2007).

From 750 to 1000 AD, much of the Lake Surprise record has poor modern analogues, though the dominance of the planktonic species *Cyclostephanos dubius* during this time (Figure 6.6), and the rapid fall in conductivity in Lake Elingamite, indicate deeper water environments and positive moisture balance across western Victoria. The decreased occurrence of severe *El Niño* events during this period, as indicated in the Moy *et al.* (2002) record, would contribute to this. Further afield, in northern Western Australia, Wohl and Fuertsch (1994) identified the occurrence of palaeoflood events during this time and the coeval occurrence of these with increased moisture in the southeast of the continent may possibly indicate a negative IOD event (warm eastern pole, cool western pole). Some support for this comes from the speleothem  $\delta^{18}\text{O}$  record from South Africa (Lee-Thorp *et al.*, 2001; Figure 6.15 (b)) which shows a significant decrease in  $\delta^{18}\text{O}$  around this time, signifying that dry conditions prevailed. A diatom-inferred lake depth reconstruction from Lake Naivasha in Kenya also indicates a major period of aridity during this time (Verschuren *et al.*, 2000), as does falling lake levels in Lake Victoria (Figure 6.15 (c)). While these records have not been correlated with IOD signals, rainfall in east Africa has been found to be closely related to positive IOD mode events (Birkett *et al.*, 1999; Saji and Yamagata, 2003; Marchant *et al.*, 2007), while negative IOD events are a major contributor to rainfall events in south-eastern Australia (Meyers *et al.*, 2007).

At this time of increased moisture in western Victoria, Rein *et al.* (2004) identify an “extreme drought” from ocean sediments off the coast of Peru and attribute this to very weak *El Niño* events. The timing of their findings does not precisely coincide with the similar findings of Moy *et al.* (2002), though this may be due to the application of a constant reservoir correction to the  $^{14}\text{C}$  dates of Rein *et al.* (2004). Within this range of dating uncertainty, Rein *et al.* (2004) correlate the inferred drought event to other periods of aridity from western USA, Patagonia, the tropical Andes, Oman, eastern Africa and other regions within the ENSO domain, demonstrating the broad-reaching nature of the event. Hodell *et al.* (1995) also found evidence of a major drought period from 800 – 1000 AD and attribute this to the collapse of Mayan civilization. The record

from Laguna Potrok Aike, east of the Andes in southern Patagonia (Figure 6.15 (d)) differs, with significant aridity not apparent until roughly 1230 AD and lasting until around 1400 AD.

Tasmanian temperatures are variable between 750 and 1000 AD. Initially, temperatures gradually decrease after 750 AD, reaching a minimum around 850 AD at a time when *Cyclostephanos dubius* and *Stephanodiscus hantzschii* dominate the Lake Surprise diatom assemblage and planktonic species again dominate the Lake Elingamite record. After 850 AD, temperatures in Tasmania begin to increase in line with increasing DI conductivity in Lake Surprise and an increase in the presence of *El Niño*-driven erosional events in Ecuador (Moy *et al.*, 2002).

The period of 750 – 1000 AD generally encapsulates a globally significant period of rapid climate change (Mayewski *et al.*, 2004). The record here though, indicates two phases to this period; one of a positive moisture balance in western Victoria with a low occurrence of *El Niño* events and cool temperatures in Tasmania from *ca.* 750 – 850 AD, and a second of an increasing number *El Niño* events with rising Tasmanian temperatures and decreasing moisture balance in western Victoria from *ca.* 850 – 975 AD. While the latter phase does not attain the significance of a lengthy drought (Figure 6.14), it does signify the beginning of a period of increased climatic variability in the records from western Victoria and elsewhere.

From *ca.* 1000 to 1400 AD there is an increased occurrence of *El Niño* events in Ecuador (Moy *et al.*, 2002), though there is a greater degree of variability within the record during this time than there was in the previous arid phase centered around 700 AD. This variability is reflected in the Lake Surprise record which shows an increased occurrence of significant ( $> 1$  SD from mean), though short-lived, drought events. However, there are also several periods ( $n = 6$ ) in which the negative deviation from the mean exceeds one standard deviation, indicating very wet conditions. Most of these periods were of a longer duration than the intervening droughts. The sediment analysis (Figure 6.13) indicates high organic and low carbonate content during these periods, which, as outlined in Section 6.4.1, is interpreted as being indicative of deep lake conditions. Conversely, the periods of elevated DI conductivity correspond with high sedimentary carbonate and low organic content, indicating lower lake levels. The

rapidity of fluctuations between these states highlights the variable nature of the climate during this time, a feature that is noted in other studies from nearby.

In Tasmania, temperatures during this period were variable, though generally elevated in relation to the entire record (Figure 6.15 (f)). In Lake Keilambete, changes reflected in the sediment composition are interpreted as being indicative of fluctuations between periods of warm and cool temperatures (Mooney, 1997). The period of change in Lake Keilambete does not correspond precisely with that identified in Lake Surprise, beginning at 1150 Cal. yr B.P. though continuing until 800 Cal. yr B.P. and thereby overlapping by approximately 200 years. The Jacka Lake ostracod-inferred salinity reconstruction (Figure 6.15 (h), Radke, 2000) has an increased sampling resolution throughout the 1000 – 1400 AD period and results mirror those in Lake Surprise, indicating considerable salinity variability on a background of generally wetter conditions.

In contrast to this, Stanley and DeDeckker (2002) identify increasingly dry and windy conditions between 1000 and 1400 AD. This may be reflecting the variable climate of the time, or may be indicating differing climatic influences between regions. The Blue Lake site of Stanley and DeDeckker (2002) lies around 800 km to the east-northeast of Lake Surprise and is located in the well documented dust-path that blows in an east and southeasterly direction from the arid and semi-arid region to the west (Figure 6.16). Over the period of increased aeolian transport to Blue Lake, Holdaway *et al.* (2002) found a lack of hearth building in the same region as the dust source and interpreted this as an abandonment of the region by the indigenous populations due to a lack of permanent water.

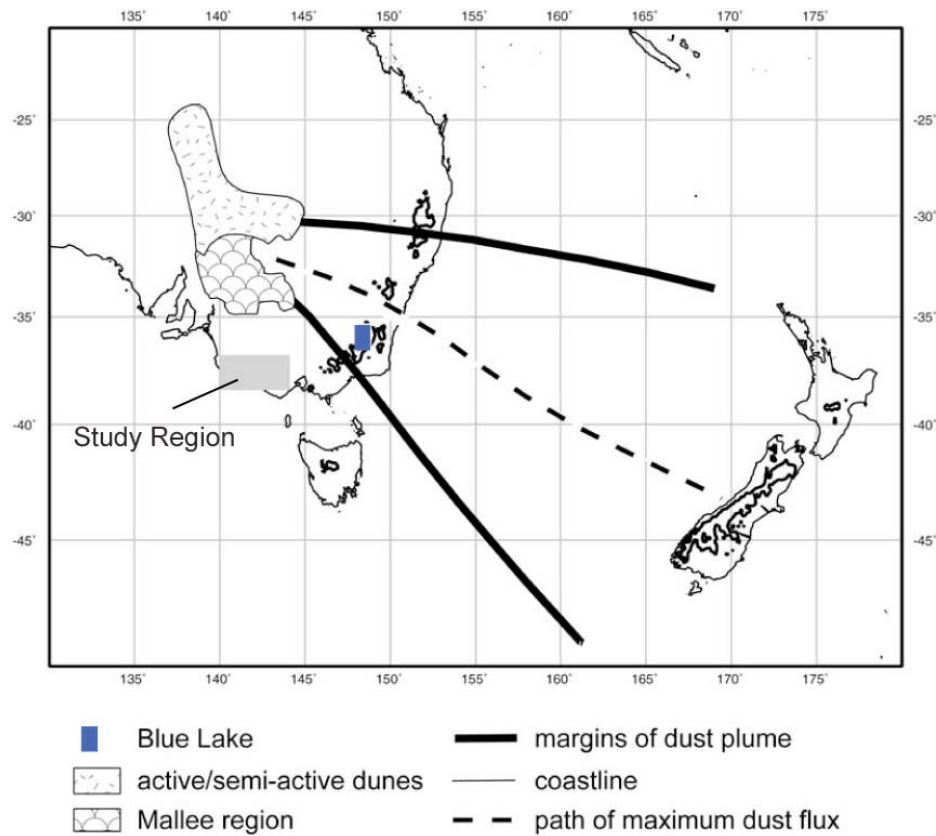


Figure 6.16: The site of Blue Lake in relation to the study region, indicating the source of aeolian sediment and the extent of the dust paths. Redrawn from Stanley and DeDecker (2002).

This suggests that the higher rainfall evident within the variable Lake Surprise record, and mirrored in the Jacka Lake record, throughout this time, may have been confined to locations nearer the coast. This could be interpreted in terms of the origin of the precipitation events being centered in the Southern Ocean, rather than in the form of northwest cloudbands emanating from the eastern Indian Ocean. Tentative support for this interpretation comes from the South Georgia record (Figure 6.15 (a)) which shows decreasing temperatures throughout this period, indicative of a northward migration of the belt of westerly winds (Rosqvist and Schuber, 2003). The South African speleothem record meanwhile (Figure 6.15 (b)), which lies at the opposite pole of the IOD to where the northwest cloudbands originate, indicates generally warm and wet conditions at this time, though an increased variability is noted. Goodwin *et al.* (2004) reconstructed the Southern Annular Mode (SAM) index from 1300 AD and identify increased frequency and amplitude of variability between 1300 and 1500 AD, indicating greater and more frequent northward migration of the mid-latitude westerlies. It is possible that this may be a source of the increased variability captured in the Lake Surprise record. If so, this

suggests that Goodwin *et al.* (2004) have captured the end of an extended period of such variability which may extend back a further 300 years.

The period of 1500 – 1850 AD is most commonly associated with the Little Ice Age (Grove, 1988), a time of glacial advances in many regions of the world and a period identified by Mayewski *et al.* (2004) as one of rapid climate change. On a centennial scale, the record from Lake Surprise has a freshening trend over this period, though on a multi-decadal time scale, there are numerous positive and negative deviations from the mean state (Figure 6.14) and a reduction in the amplitude of variability. A similar pattern is evident in the data of Moy *et al.* (2002) which indicates reduced amplitude of sub-decadal variability, on a background of gradually decreasing occurrences of *El Niño* events at a centennial scale. The rapid decline in the presence of aeolian grains in the Blue Lake record indicates the broad spatial signature of this wetter climate regime in western Victoria. Apart from a slight increase in some pollen types and fern spores, there is scant evidence of this in the Lake Keilambete record though (Mooney, 1997). However, interpretation of this period is hampered by the slow sedimentation rate during this section of the record (Mooney, 1997).

The SAM reconstruction of Goodwin *et al.* (2004) shows a pattern of slower variations between 1500 and *ca.* 1850 AD and is weakly positive, indicating increasing wind strength over southeast Antarctica. This pattern is also reflected in west Antarctica, signifying meridional changes in the atmospheric circulation over the southern Pacific Ocean (Kreutz *et al.*, 1997). Temperatures in South Georgia were cold (Figure 6.15 (a)), though more so towards the latter half of this period. In southern Patagonia, lake levels rose considerably (Figure 6.15 (d)) on the background of a cool, wet climate (Haberzettl *et al.*, 2005).

Temperatures in Tasmania show a gradually cooling climate after 1500 AD with reduced multi-decadal variability, possibly due to the influence of strengthening westerly wind patterns at this time (Shulmeister *et al.*, 2004). Similarly, an increased south-westerly circulation pattern over New Zealand led to a period of glacial advance (Shulmeister *et al.*, 2004), with maximum glaciation between 1725 and 1740 AD (Winkler, 2000; 2004), though the timing of individual advances differs regionally. North Island speleothem  $\delta^{18}\text{O}$  records also indicate falling temperatures around this time (Williams *et al.*, 2004).

Eastern equatorial Africa displays regionally complex rainfall variation during this time. In the east of the region the reconstructions from Lake Naivasha (Verschuren *et al.*, 2000) and Lake Victoria (Figure 6.15 (d); Stager *et al.* 2003) and ice cores from Mount Kenya (Karlen *et al.*, 1999) and nearby Mount Kilimanjaro in Tanzania (Thompson *et al.*, 2002), show that the climate was variable, though generally wet. Towards the west of equatorial east Africa, opposing evidence emerges with periods of drought coincident with the periods of increased rainfall in the east (Russell *et al.*, 2007). The underlying causes of this east/west rainfall gradient remain unresolved (Russell *et al.*, 2007). In the south of the continent in South Africa, the climate was cool and dry, reaching a temperature/precipitation minimum at *ca.* 1750 AD (Figure 6.15 (b)), marking the coldest and driest period of the last two millennia. This coincides with the timing of the coolest sea surface temperatures (SSTs) identified in a ~ 300 year reconstruction of IOD-driven SST change from Madagascar corals (Zinke *et al.*, 2004).

It appears therefore, that the wet climate and deep lake levels in western Victoria may have been driven by an increased westerly circulation at this latitude, bringing moist maritime air masses into the study region and beyond. As such, average annual temperatures would have decreased which, in addition to the low frequency of *El Niño* events, would have contributed to lower evaporation rates and the maintenance of deep lake environments. Aridity in most of east Africa suggests that the IOD may have tended towards the negative mode (cool west - warm east). A warm eastern Indian Ocean would also have contributed to rainfall across the study region, especially if this was associated with increasing *La Niña* events (Meyers *et al.*, 2007).

In a model of historical lake-levels from three lakes in western Victoria, Jones *et al.* (2001) calculated that, prior to 1840 AD, a P/E ratio of between 0.94 and 0.96 existed in the region. Historical data indicates that lakes in the study region were full and, in some cases, overflowing (Currey, 1970). However, Jones *et al.* (2001) identify a large shift in precipitation and/or evaporation occurred between 1840 and 1863 AD, reducing the P/E ratio to 0.79 and marking the end of what the results of this study indicate, was an extended period of elevated rainfall. The purported shift in P/E ratio does not exactly correspond to changes in Lake Surprise, which records increasing conductivity around 1875 AD, though the impacts of this shift are clearly evident.

As demonstrated above, the DI conductivity record of Lake Surprise reflects the PDSI for the southeast of mainland Australia and records the occurrence of several historical drought periods. The palaeoclimate record presented here places the climate of the last century in a long-term perspective and indicates that the 20th century climate differs markedly from that of the preceding four and a half centuries. These, in turn, differ markedly from the centuries preceding them. The changes evident in the record of the last *ca.* 1500 years represent one of the first fine-resolution climate reconstructions from mainland Australia to capture the periods commonly referred to as the Medieval Warm Period and the Little Ice Age.

### **6.8 The Medieval Warm Period and Little Ice Age in western Victoria**

Prior to the development of the palaeoclimatic reconstruction presented in this study, there has been scant evaluation of either the LIA or MWP in Australia. Some studies (e.g., Mooney, 1997; Radke, 2000; Stanley and DeDeckker, 2002) provide an indication of climatic shifts during these time periods, though the generally low sampling resolution of these studies precludes more definitive interpretations. In contrast, Hendy *et al.* (2002a) provide a highly resolved record of palaeosalinity and palaeotemperature derived from Great Barrier Reef coral, though this only extends back to *ca.* 1580 AD and thereby excludes the MWP. Even so, these studies add to a growing body of knowledge about the regional climate and its influences during these periods and, in conjunction with this record, can provide an illustration of differing palaeoenvironments.

As discussed in Chapter 2, there is still debate as to whether the MWP was a global-scale event. Primarily, this is due to a lack of temporal and spatial coherence in the expression of the MWP, which commonly takes the form of a temperature or hydrological anomaly. As outlined by Bradley *et al.* (2003a), in terms of temporal cohesion, the argument is confounded by many researchers identifying a “Medieval Warm Period” for any climatic anomaly that occurs during the historical Medieval period (500 – 1500 AD). For example, Abrantes *et al.* (2003) claim decreased river discharge off Portugal between 550 and 1300 AD as evidence of a MWP that lasted 750 years. Other studies are more selective in their boundaries of the MWP, though occasionally at the expense of considering other climate anomalies as being an aspect of MWP climate. For example, Laird *et al.* (1998) cite a period of aridity from 1000-

1200 AD as being the manifestation of the MWP in the northern Great Plains of the United States, though do not consider a similarly arid period between 700-850 AD to be part of the MWP. Similarly, from a study in California, Li *et al.* (2000) state that a dry climate from 950-1220 AD corresponds with the MWP, though a wet period from 1220-1480 AD is not considered to be an aspect of MWP climate. Interestingly, this ‘non-MWP’ humid period of 1220-1480 AD closely corresponds with a period of aridity in southern Patagonia from 1230-1410 AD (Haberzettl *et al.*, 2005:283) which is claimed by those authors to be representative of “a signal of the Medieval Climate Anomaly in south-eastern Patagonia” (Haberzettl *et al.*, 2005:283).

The assigning of boundaries to the MWP therefore seems arbitrary and appears to depend upon when an ‘appropriate’ climatic anomaly is identified. Yet focusing on a certain time period (e.g., 1000-1200 AD), or on a desired climatic anomaly (e.g., aridity or elevated temperature), is to the detriment of a thorough examination of the climate processes that have driven synchronous, though contrasting, climates between regions. In order to promote a clearer understanding of intra- and inter-regional climates around the turn of the first millennium, it is apparent that there is a need for a defined chronozone to be assigned to the MWP (*sensu* Björck *et al.*, 1998).

Given the lack of clear temporal boundaries for this period then, it is difficult to definitively state that any anomaly identified in the lakes studied here is an expression of the MWP in Australia. It is clear however, that significant climate anomalies were present towards the end of the first millennium and the beginning of the second. These climate anomalies take the form of a multi-decadal drought occurring between *ca.* 680 and 780 AD and a period of marked climatic variability between *ca.* 1000 and 1400 AD. The intervening 200 year period is characterised by what was probably a wet century from 800 – 900 AD, followed by a century where aridity was gradually increasing.

Within the broad ‘Medieval’ period then, there are differing climate signals – very dry, probably wet, increasingly dry and highly variable – rather than one clear event. The climate of the late first millennium to early second millennium in western Victoria can therefore be considered to have been highly variable, a finding reflected in the Lake Keilambete record (Mooney, 1997). Indeed, the climate across this time period was the most variable in the last 1500 years. This variability appears to be driven by the major circulation systems of ENSO and the IOD. The underlying cause of variability within



these systems though, is yet to be fully determined. In terms of ENSO, millennial-scale variability may be driven by internal dynamics of the ENSO system (Moy *et al.*, 2002; Cobb *et al.*, 2003). The relatively new discovery of the IOD and lack of long-term IOD reconstructions means that little is known about drivers of variability of this system and, as discussed above, debate remains as to the degree that the IOD is independent from ENSO.

In comparison to the MWP, the Little Ice Age is better constrained with the majority of studies identifying a climatic anomaly generally occurring between 1400 and 1850 AD, though there remains debate around the suitability of the term and its application to varying climatic anomalies (e.g., Landsberg, 1985; Jones and Briffa, 2001; Mayewski *et al.*, 2004). From the records presented here, the period of *ca.* 1500 to 1850 AD can be characterised as being wet and probably cool on a centennial scale. At decadal-scales, there is a reduced amplitude of variability compared to the remainder of the Lake Surprise record and, while some arid periods are evident, none compare in either severity or duration to the present drought. From *ca.* 1700 – 1875 though, the evidence indicates a persistently wet climate. A reduction in dust transport and deposition (Stanley and DeDeckker, 2002) from locations beyond the study region suggests that these wet conditions may have extended beyond the boundaries of the study region. As was the case in the preceding period, the climate at the time appears to be a manifestation of ENSO and IOD influences, though the effects of the westerlies also appear to be more important during this time (Shulmeister *et al.*, 2004).

This interpretation of the mechanisms behind the identified climate changes during the periods associated with the so-called Medieval Warm Period and Little Ice Age is seemingly at odds with other interpretations. Specifically, Turney and Palmer (2007) propose that the MWP (defined by them as 950 – 1350 AD) could be characterised as persistent *La Niña* conditions, while the LIA (defined as 1350 – 1850 AD) may have been dominated by *El Niños*. This would lead to climate conditions in southeast Australia that were the opposite to those identified in this study and also contradict the findings of Moy *et al.* (2002). In addition, Turney and Palmer (2007) cite evidence of warm SSTs in Great Barrier Reef coral during the LIA (Hendy *et al.*, 2002a) as supporting their claim. However, a warm western Pacific is indicative of *La Niña* conditions (not *El Niño* conditions as inferred by Turney and Palmer (2007)) and thus lends support to the interpretation proposed here.

In conclusion, there is clear evidence from both Lake Elingamite and Lake Surprise indicating a highly variable climate between *ca.* 680 and 1400 AD, with a high frequency of drought episodes. Whether this period, or a portion of this period, represents the manifestation of the MWP in western Victoria is unable to be determined due to the fluidity of the temporal boundaries commonly assigned to the MWP. From *ca.* 1500 and 1850 AD, the evidence indicates a lengthy wet and probably cool period. Due to the generally better constrained temporally boundaries of the LIA, it could be concluded that this extended period of cool wet climate is the manifestation of the LIA in western Victoria.

The discrepancies between the interpretation of results in this study and that of Turney and Palmer (2007) for instance, highlights the clear need for additional research from the Southern Hemisphere in general (Jansen *et al.*, 2007), and the Australasian region in particular (Turney *et al.*, 2006), into fine-resolution climate records in order that the underlying processes that are driving observed changes can be more accurately determined.

## CHAPTER 7 – CONCLUSIONS

This thesis has presented the first high-resolution quantitative evidence of climatic variability in western Victoria over the past *ca.* 1500 years by studying sensitive sites and utilising a sensitive climate proxy. The finding of a replicated climate signal in two geographically separated lakes within the same climatic region, provides evidence of a regional-scale climatic signal which has been shown to have fluctuated considerably over time. Due to unforeseen challenges, such as the need to develop a new diatom-conductivity transfer function tailored specifically for application to low conductivity lakes, the plan and initial aims of the project necessarily changed. However, the focus of the project; reconstructing the short-term climate history of southeast Australia in fine-resolution, remained unchanged. The following discussion concludes this thesis by reviewing and assessing the varying components of the project, summarising the research findings and identifying implications for further research.

### 7.1 Assessment of the key components of the project

#### 7.1.1 Site selection

As with all palaeoenvironmental studies, specific sites for this study were chosen based on their ability to satisfy the aims of the project. As discussed in Chapter 4, rigorous site-selection criteria were applied with many potential sites being rejected. The two sites chosen, Lakes Elingamite and Surprise, proved to be effective in recording a high-resolution history of climatic change, though their individual responses to climate differed.

The choice of two lakes that lay distant to each other at a local scale proved instructive and demonstrated that the climate signal being identified was a region-scale signal and not one of local origin. The records were complementary in a number of ways. For instance, the maintenance of a climate signal in Lake Surprise during historical times enabled a correlation to be made with instrumental records and proved that the lake was climatically sensitive. Due to a greater degree of post-European disturbance, this was not possible with Lake Elingamite. However, the coeval changes evident in the lakes in the pre-European record demonstrated that Lake Elingamite was also responding to climate changes. In addition, the Lake Elingamite fossil diatom samples had good analogues in the calibration set, providing support to the interpretation of sections of the Lake Surprise record in which there were poor analogues in the modern data set.

In summary, the application of strict selection criteria successfully identified lakes that could satisfy the aims of the project. While it is perhaps unfortunate that Lake Elingamite demonstrated a more muted response to climate forcing, this did provide a stark illustration of the severity of the late first millennium drought. The degree of detail evident in the Lake Surprise record proved highly informative and provided detail of processes such as mixing regimes, species turnover and even short-lived episodes of turbidity, which enhanced the resulting interpretation.

### 7.1.2 Transfer function development and application

The sites identified as most appropriate for this study had low conductivity values and preliminary reconstructions from Lake Elingamite, derived from the application of an existing transfer function (Gell, 1997) had large predictive errors. In part this was due to the long salinity gradient of the data set and the low salinity of Lake Elingamite. The need therefore arose to develop a diatom-conductivity transfer function for accurately reconstructing conductivity from low salinity sites. Initial sampling of modern diatom assemblages was hampered by the dramatic loss of water resources during the current drought, though this was overcome by broadening the geographical range of sample sites and sourcing samples from an external agency. However, the externally sourced samples had a less extensive suite of measured environmental variables and measurements of potentially important variables, such as ionic chemistry for instance, were absent. Once both training sets were merged, Canonical Correspondence Analysis identified conductivity as the variable exerting the greatest and significant influence on diatom taxon variance, tested using 9999 random Monte Carlo permutations.

The conductivity gradient of the modern dataset was intentionally short (81.3 – 21380  $\mu\text{S}/\text{cm}$ ) but comparable to several other salinity transfer functions from freshwater systems (e.g., Ng and King, 1999; Ryves *et al.*, 2002; Tibby and Reid, 2004). The effect that differing diatom count sizes had on the jack-knifed  $r^2$  and RMSEP was assessed and a count size of 200 valves was identified as having the lowest errors and highest jack-knifed  $r^2$ . Using the ‘200-dataset’, a total of 169 taxa were recorded from the 47 sites that constituted the final modern calibration set.

In order to develop a model with the lowest errors and highest  $r^2$  values, rare taxa were removed. Varying definitions of 'rare taxa' were applied and each resulting dataset was evaluated by comparing errors and  $r^2$  values after employing three different cross-validated (jack-knifed) regression techniques (WA-Inverse, WA-Classical and WA-PLS) with outliers removed. In order to identify the model with the highest accuracy at low conductivity, the five best performing models (based on jack-knifed  $r^2$  and RMSEP values) were ranked in order of the average of their residuals that were less than the observed conductivity of Lake Elingamite. The chosen model has a jack-knifed  $r^2$  of 0.89 and an RMSEP of 0.238 log  $\mu\text{S}/\text{cm}$ . As a percentage of the gradient length, the RMSEP was 9.8%. The model compares favourably with other diatom-conductivity transfer functions.

The main weakness of the transfer function was the inability to reconstruct ionic chemistry and determine the degree to which this may have influenced diatom species distribution. As outlined above, this was due to a lack of suitable calibration sites due to the impact of the present drought, which is ironic, given the aims of the project.

To conclude, the aim of this portion of the project was to develop a diatom-conductivity transfer function that was tailored specifically for application to fresh systems. In the process, every effort was made to develop the most robust diatom-conductivity transfer function possible and minimise reconstructive errors. The measures of model performance (jack-knife  $r^2$ , RMSEP and RMSEP as a percentage of gradient length) indicate that the desired aim was achieved.

### 7.1.3 Reconstruction of conductivity

Based on an evaluation of counting efficiency (Pappas and Stoermer, 1996), fossil counts were limited to 200 valves per sample. Using the model discussed above, palaeoconductivity reconstructions were performed on all three cores (Figures 5.40, 5.41 and 5.42). Some portions of the reconstructions had large predictive errors, though these were generally restricted to samples with a reconstructed conductivity greater than 2500  $\mu\text{S}/\text{cm}$ . Below this conductivity, samples generally have a very narrow range of reconstructive error, highlighting the effectiveness of the transfer function from reconstructing very low salinities.

The conductivity reconstruction was evaluated using the Modern Analogue Technique (MAT). Fossil samples with a squared-chord distance greater than 10<sup>th</sup> percentile of the dissimilarity coefficients of the calibration set were considered to have poor modern analogues. The assignment of such a criteria is comparable to other studies (e.g., Laird *et al.*, 1998; Kauppila and Valpola, 2003; Tibby *et al.*, 2003; Bennion *et al.*, 2004) and stricter than some (e.g., Bradshaw and Anderson, 2001; Ryves *et al.*, 2002). The transfer function was validated by comparing the recent Lake Surprise reconstruction and instrumental records, specifically, the Palmer Drought Severity Index (PDSI) as calculated by Ummenhofer *et al.* (2009). These two measures compared well, though errors the dating of the most recent sediments consistently offset the reconstruction by approximately six years.

The majority of fossil samples had good or very good modern analogues in the calibrations set, though 44% of core LST1 was determined to have poor modern analogues. Much of this was due to the absence of two key fossil species from the modern dataset; *Cyclostephanos dubius* and *Stephanodiscus hantzschii*. In sections where these species dominated the diatom assemblage, reconstructions were unreliable. To a degree, this was overcome by referring to the Lake Elingamite reconstruction for an indication of climate over the relevant periods, however, the addition of these species to the calibration set would prove instructive for interpretation of the longer record (Figure 3.19; Aline Philibert, unpublished data) or if further work on Lake Surprise is undertaken. In the absence of modern sites containing these species, the option of importing data from international diatom databases (e.g., the European Diatom Database – EDDI) could be explored. While this is not an ideal solution, it has previously been successfully employed where similar problems exist (e.g., Mills, 2009). In the case of this project though, this option was not explored as the record of Lake Elingamite, which had good, and very good, modern analogues throughout, provided support for interpretation of the affected portions of the record.

Developing an accurate chronology for the Lake Surprise and Lake Elingamite records posed similar dating challenges as has been experienced in other lakes of the region (e.g., Mooney, 1997). Every effort was made to overcome the identified issues and, where possible, evidence from other proxies was incorporated. While it is acknowledged that the final chronology from both lakes is not ideal, it is considered to be the best that could be achieved given the available data. Justifications for the final

chronologies are provided in the body of the thesis. Application of the proposed chronologies reveal coincident changes within the two study lakes which also correspond to changes recorded elsewhere, suggesting that the mechanisms responsible for the observed changes may be operating extra-regionally.

## **7.2 The effectiveness of diatoms as indicators of climate change**

The paucity of short-term, high-resolution, records from the Australasian region has been long recognised (e.g., Wasson, 1990) and recently reiterated (e.g., Turney *et al.*, 2006). In part, this can be attributed to the fact that, compared to other regions of the world, fewer opportunities exist in Australia for developing high-resolution palaeoclimate records. For instance, long term tree-ring records are rare due to a dearth of very long-lived species, there are no permanent ice fields to provide ice-core records and no known varved lakes, while coral and speleothem studies are yet to provide any continual records of the last two millennia from the region. Palaeolimnological research therefore represents one of the few opportunities to provide continuous, decadal to sub-decadal resolution data over a long-term period.

Several lake-based proxies are amenable to such high-resolution analysis and, for reasons outlined in the body of the thesis, diatoms were chosen for this study. The sensitivity of diatoms to chemical parameters is well documented and they have been successfully used for palaeoclimate reconstructions in the past (e.g., Fritz, 1990; Fritz *et al.*, 1994; Gasse *et al.*, 1997; Laird *et al.*, 1996a, 1996b, 1998; Gell, 1998). In this study, they once again proved effective in reconstructing climatic variations, with fluctuations of inferred conductivity over historical times corresponding well to the PDSI drought index for the region. Furthermore, the interpretation of the results was aided, and strengthened, by previous studies of the ecology of key species such as *Cyclostephanos dubius*. Importantly though, the short life span of diatoms, in conjunction with their sensitivity to chemical parameters, ensures that they respond quickly to changing conditions, making them ideal for high-resolution records. Evidence of this responsiveness is apparent in the rapid and dramatic shifts in diatom species composition in the Lake Surprise record. This provides a level of detail that few other biological indicators could equal.

However, limitations exist in terms of interpreting the shifts evident in the diatom data. In terms of this study, these limitations were due to factors including poor modern analogues in the training set, the inability to determine changing ionic chemistry, the broad ecological tolerances of some key species, an absence of further proxies to provide complementary or contrasting data, and a lack of climate records from the region with which to compare the interpretation.

Despite these limitations though, diatoms proved effective as a means to reconstruct changing moisture balance in the study region. From an Australian palaeoclimate perspective, diatoms are one of the few proxies that can achieve this at highly resolved scales. The only caveat is that this is only achievable where suitable sites exist.

### **7.3 Summary of findings**

The diatom-inferred conductivity reconstruction from Lake Surprise corresponded well with the Palmer Drought Severity Index (PDSI) over the historical period and thus enabled a record of drought severity, frequency and duration of the last *ca.* 1500 years to be developed. Comparisons of reconstructions against results from other relevant literature showed a good correspondence. In particular, the reconstruction of major *El Niño* events from Ecuador (Moy *et al.*, 2002) was clearly reflected in the results from the lakes in western Victoria.

Droughts, defined as having a reconstructed conductivity more than one standard deviation greater than the mean, were more severe between *ca.* 700 and 1200 AD. In particular, the period of *ca.* 680 – 770 AD was extremely dry, with increased conductivity evident in both lakes. This period corresponds with the maximal number of *El Niño* events per century of the last 2000 years (Moy *et al.*, 2002). Thereafter, from *ca.* 900 to 1400 AD, the climate of western Victoria was highly variable with the DI conductivity record indicating droughts of short duration, interspersed with longer periods of positive moisture balance. Comparisons with other studies suggest that these drought periods were driven by an increasing frequency of major *El Niño* events. However, based on South African speleothem results (Lee-Thorp *et al.*, 2001), it is proposed that the Indian Ocean dipole (IOD) may also exert an influence on the patterns of aridity.



From *ca.* 1625 to 1880 AD, the climate was wet. Lake Elingamite probably overflowed during this time, while Lake Surprise had a freshening trend. The occurrence of major *El Niño* events during this time was low and temperatures in Tasmania had reduced decadal variability (Cook *et al.*, 2000). South African climate was slowly drying and becoming cooler during this time, and it is proposed that this may reflect an increased occurrence of a negative IOD, bringing warm sea-surface temperatures to the east of the Indian Ocean and hence, contributing to rainfall over the region (Meyers *et al.*, 2007).

The current drought is placed into stark historical perspective by these records. When considered in terms of both the duration and intensity of the drought, there has not been a comparable event for *ca.* 1200 years. There have been more severe droughts in the intervening period, though none were of a similar duration to the present drought. In addition, the evidence presented indicates that agricultural exploration and settlement in the study region occurred towards the end of what had been an unprecedentedly long wet phase, perhaps providing an inaccurate representation of the natural state of the landscape.

In terms of a critical assessment of the reconstruction, portions of the Lake Surprise record suffer from poor modern analogues which, to a degree, hinders the interpretation. By and large though, this has been overcome through interpretation of sedimentary data, knowledge of the autecology of species and comparisons against the Lake Elingamite record. The chronology from both lakes precludes an accurate determination of the precise length of drought periods and, if further research were to be undertaken on either lake, a fine-resolution chronology would be beneficial. Finally, the length of the records, ~1500 years, precludes analysis of long-term low frequency variability. Again, this may be a worthwhile focus of further study.

Overall, the interpretation presented from the Lake Elingamite and Lake Surprise records shows a degree of climatic variability not previously seen from mainland Australia. At multi-decadal scales, the climate was significantly more variable prior to *ca.* 1300 AD with a hitherto unrecorded period of prolonged and severe aridity. At a centennial-scale, there is a *ca.* 300 year period of persistent dryness (though without the presence of major aridity) from around 1350 to 1625 AD, followed by *ca.* 250 year period (~1625 – 1875 AD) of almost consistently wet conditions prior to, and incorporating, European colonisation of the region. The current drought can now be

viewed within an historical context greater than 120 years for the first time and has been identified as a rare, one-in-1200-year event. In addition, the climatic influence of ENSO, and probably the IOD, have been recorded.

#### **7.4 Implications for further research**

While this study has provided answers to a number of questions, it has also raised issues that require further investigation. The effect of the IOD on the climate of south-eastern Australia has been documented by Meyers *et al.* (2007) and Ummenhofer *et al.* (2009) among others. However, millennial-length reconstructions of this relatively newly discovered dipole, which would have proved invaluable to the interpretation of the records presented here, are not presently available. Correlation of records from South Africa (such as the speleothem record of Lee-Thorp *et al.* (2001) for example) to the IOD may be a valid line of inquiry to achieve this. This may also assist in determining a long-term history of the relationship between the IOD and ENSO, which would provide invaluable knowledge towards understanding the long-term climatic variability of south-eastern Australia.

One aspect that is absent from this study is a clear signal of temperature change. While it has been clearly demonstrated that the P:E ratio has fluctuated over time, which precise component of that ratio changed during periods of positive or negative moisture balance has only been hypothesised. Therefore, whether the climate was warmer at the time of increased drought frequency for example, remains unresolved. In this regard, there is a priority to complement studies such as this with those from sites with complete ostracod records in order that palaeotemperature evidence (Chivas *et al.*, 1986) can be matched with the P:E reconstruction.

Lake Surprise has been shown to be a highly sensitive site, capable of recording high frequency climatic fluctuations in the fossil diatom data at least, with a sedimentation rate amenable to high-resolution studies. Coring of the complete sedimentary sequence has demonstrated that the record continues beyond the last glacial maximum to approximately 30,000 cal yr B.P. (Builth *et al.*, 2008). Therefore, the possibility exists to explore numerous short- or long-term climate events and periods of interest in high resolution, that have either not been clearly documented previously or for which the short-term variability within the periods is unknown. Examples include, the so-called

8.2 ka event, the Antarctic cold reversal, the Younger Dryas, the Holocene/Pleistocene boundary and the presence/absence of 'Bond cycles' or 'Heinrich events' in Australian palaeoenvironmental history. In addition, the extension of the drought history presented here would provide a better understanding of long-term, low frequency climatic variability. Any further work undertaken on this lake though, would benefit from a high-resolution dating regime to account for changing sedimentation rates over time and provide precise information regarding the duration of identified events. This would also need to be corrected following an accurate determination of the contribution of old carbon to the lake system. In addition, detailed geochemical analysis of the carbonate laminations may prove to be very insightful. The recent acquisition of Australia's first Itrax scanner, which can record elemental variations in a sediment core at very fine (200  $\mu\text{m}$ ) sampling resolution (Croudace *et al.*, 2006), will prove very helpful in this regard, and in terms of other fine-resolution analyses.

Finally, given the dearth of fine-resolution studies from Australia, and considering the evidence presented here of major fluctuations in effective moisture in the southeast of the continent during the past 1500 years, there is a clear need for further research into the climate of the last two millennia. On a continental scale, a broader spatial distribution of studies could inform the relative influence of the major atmospheric and oceanic influences driving the identified fluctuations in effective moisture. On a regional scale, further studies from south-eastern Australia are required in order to determine if, and how, other sites responded to the changes identified. The presence of several studies from the region could address questions about the replication of findings, the responsiveness of individual sites, the coherence between records and the manner to which this may vary as a function of the lakes' position in the landscape (*sensu* Webster *et al.*, 1996).

## **7.5 Conclusion**

Mid-way through this project, in 2006, political leaders in Australia were informed that the drought afflicting regions of Australia was the worst in a thousand years (Shanahan and Warren, 2006). The results of this study provide the first evidence that this statement is only partly correct and that, in fact, the statement is an underestimation of the true context of the present drought. Placing extreme events into an historical perspective is a strength of palaeoclimatic research and there remains a clear need for a

greater focus on fine-resolution research from Australia into the climate of the past two millennia. The formation of the Aus2k working group (see Aus2k – website), as a component of the PAGES focus 2 (regional climate dynamics) ‘2k Network’, is an exciting step towards addressing the paucity of such research.

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