A 2000 year record of environmental change from Tocal Homestead Lagoon, eastern Australia

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I, Duncan Edward Cook, declare that this thesis, submitted in fulfilment of the requirements for the award of Doctor of Philosophy, in the School of Geosciences, The University of Sydney, has not been submitted for a higher degree at any other academic institution and, unless otherwise acknowledged, is my own work.

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Duncan Edward Cook

Tocal Homestead Lagoon

Abstract

A high-resolution sedimentary record from Tocal Homestead Lagoon, in the lower Hunter valley of eastern Australia, has provided detailed information on environmental change during the last *c.* 2000 years. Twenty-four cores were retrieved from the lake. These have been correlated using magnetic susceptibility measurements, allowing the construction of a basin-wide magneto-stratigraphy. The core with the longest, most complete and the highest resolution record (core TCA9b) was sub-sampled and analysed for a suite of physical, magnetic and chemical properties. The lake sediments have been precisely and accurately dated using a range of independent methods providing arguably the most detailed, continuous record of late Holocene environmental change on the Australian continent. Radiocarbon, palaeomagnetic and radiometric dating have been used to identify the sedimentary horizon of first European contact in the early 19th century AD. This has allowed a detailed record of catchment-wide soil loss for the period *c.* 1820 to 1998 AD to be obtained, providing a rare opportunity to examine the timing and magnitude of the landscape disturbance that accompanied European settlement in the 19th century. In addition, insights into the state of the pre-impact environment during the last *c.* 2000 years have also been gained, providing much needed information on late Holocene environmental change and pre-European human impacts.

The Tocal Homestead Lagoon record has shown that periods of contrasting climatic conditions prevailed during the late Holocene in eastern Australia. These environmental shifts have, so far, not been recorded in any of the pollen-based palaeoenvironmental records from the region. In particular, the Tocal record has provided some evidence of a prolonged warm period after *c.* 900 AD, broadly coincident with the Medieval Warm Period. The lake sediments have also supplied indirect evidence of the chronology of Aboriginal occupation at the site. Burning of the immediate catchment has taken place regularly since *c.* 150 BC. This is in agreement with most fossil charcoal records in the region, which show that regular firing of the environment has occurred during the last several millennia. More explicit evidence of Aboriginal occupation comes from the lake sediment chemistry, with the concentration of P increasing above background levels from *c.* 800 AD, roughly 1000 years before European settlement in the region. By the time the first European had entered the catchment, P levels had already doubled above the site's pre-800 AD, background level. In the absence of any other plausible explanation, this has been cautiously interpreted as a record of pre-European human activity within the Tocal catchment.

The mean rate of catchment-wide soil loss for the *c.* 2000 years before the arrival of Europeans was 18 $t\,km^{-2}a^{-1}$. This may be contrasted with the mean rate of erosion experienced under European land use of 228 $t\,km^{-2}a^{-1}$, an increase of more than an order of magnitude. Although this mean rate for the post-settlement period is comparable to modern rates of catchment sediment yield in eastern Australia, it masks periods of much greater denudation.

The arrival of Europeans at Tocal is clearly recorded in the lake sediments as a pulse of topsoil-dominated material, which marked a spectacular rise in erosion to 638 t $km^{-2}a^{-1}$, more than 200 times higher than the rate at which soils were being eroded in the preceding four centuries. Truncated A-horizons and compacted soil profiles in the catchment today provide modern records of this past denudation. The catchment's response to the arrival of Europeans, the clearing of land for farming and the impact of thousands of hard-hoofed animals in the catchment was both swift and intense. In contrast, the pollen record for the period immediately before and after European settlement shows little variation, which has important implications for our understanding of vegetation response to human impacts in the Australian environment.

Although the initial shock of European impact resulted in a dramatic increase in erosion, the greatest rates of catchment-wide soil loss at Tocal took place during periods of intense cattle ranching at the beginning of the 20th century. Earlier, when agriculture at the site was crop-dominated, the catchment appears to have been more stable and erosion less severe. Lower rates of erosion have been recorded in the last decade at Tocal, which may represent the impact of more sustainable farming practices. This pattern is somewhat different to that recorded in other long-term sedimentation records in eastern Australia, which show that the greatest erosion rates immediately followed European settlement.

Decadal-scale climatic variations appear to have had little direct influence on the history of erosion, although there is good evidence that extreme climatic conditions may have been responsible for amplifying the already severe impacts of intensive land use. This may be in part due to the dominant role of land use in controlling erosion in the historic period. However, the observed cycles of drought and flood in the region in the decades before European settlement in *c.* 1820 appear to have had no impact on soil erosion either. It appears that the pre-European environment may have had some level of resilience to short-term climatic fluctuations and Aboriginal land use. If some level of equilibrium existed in the pre-European environment, then it was both easily and rapidly disturbed and any resilience such systems may have had to minor climatic fluctuations appears to have been incapable of coping with the environmental changes that accompanied the arrival of Europeans in Australia.

Acknowledgements

Very many people worldwide need to be thanked for their help with this research. First, I am very grateful to my supervisor at The University of Sydney, Dr Stephen Gale, who first suggested examining the site of Tocal for palaeoenvironmental information. Throughout my candidature he has been a dedicated and conscientious supervisor, providing excellent guidance with the research, detailed comments and corrections to my writing. The essential ability to maintain a sense of humour throughout it all was greatly appreciated. His help in circumnavigating both theoretical and practical obstacles in particular has been invaluable. I am also indebted to him for his instruction and assistance in the field and laboratory, as well as for securing funding to support both of these. Funding for this research was kindly provided by the Australian Institute of Nuclear Science and Engineering (AINSE), the Australian Research Council (ARC), the Quaternary Research Association (QRA), the Smithsonian Institution and The University of Sydney. This research was undertaken while I was in receipt of a Henry Bertie and Florence Mabel Gritton Postgraduate Research scholarship and tuition waiver scholarship from The University of Sydney, for which I (and my landlord) am extremely appreciative.

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Chapter One

Introduction

1.1 The Hunter valley

The Hunter valley of central eastern Australia has the dubious distinction of having one of the longest records of environmental degradation on the continent. Following the first observations of its coal-rich sea cliffs at the end of the 18th century by Europeans, land use in the region followed a course of continued exploitation of its resources. Mining of coal for use in the new settlement of Sydney and for export originally brought European colonists to the coastal fringe of the Hunter valley. Following two decades of quarantine, during which time the settlement of Newcastle was used as a penal colony, the discovery of rich supplies of red cedar on the banks of the Hunter River drove exploration and settlement inland. The discovery of rich alluvial plains (increasingly cleared by timber getters) led to the rapid onset of farming in the 1820s, with the Hunter valley quickly becoming a major source of food for the new colony. Today roughly 40% of the land in the valley is utilised for agriculture (Australian Bureau of Statistics, 2003).

There were early signs that the intensive use of land in the region was taking its toll; some commentators in the 19th century even suggested that the silted-up rivers and increasingly large gullies were somehow connected to human activity. This may in part explain why the Hunter valley was one of the first regions in Australia in which soil erosion was monitored. The first scientific records of looming environmental degradation date from 1942 and showed that up to 36% of the region was gullied and only 42% of the 25 000 km^2 area remained unaffected by erosion (Kaleski, 1945). As in much of Australia, soil erosion continues to be a major environmental problem: by the late 1980s, over 80% of the region was affected by erosion (Emery, 1989). A famous and often cited study from the Hunter valley's vineyards produced the startling finding that the production of a single bottle of wine resulted in the erosion of an equivalent volume of soil (Loughran *et al.*, 2000). In many ways, the region is typical of farmland across temperate Australia, and the findings from the Hunter vineyards have received widespread media attention and have raised the spectre of potentially irreversible soil loss in the Australian environment to the general public.

1.2 Soil erosion

Management of the accelerated loss of soil from land across Australia is perhaps the greatest challenge that faces agriculturalists. Australia's ongoing dependency on its primary production sectors and, by extension, on its soil, adds both economic and social dimensions to the problem of soil loss. Even 30 years ago, in 1975, the estimated cost of repairing lands already degraded had reached $675 M (Anon., 1978). Much scholarship in Australia has been dedicated to the concept of tolerable soil loss and to the question of how much soil loss is acceptable (see, for example, Edwards [1988]). Perseverance until recently with the philosophy of 'soil loss tolerance' in agriculture, where the maximum rate of erosion that will still permit crop productivity to be maintained (Beckmann and Coventry, 1987), has undoubtedly worsened the erosion problem. Australian soils tend to be shallower than those found elsewhere in the world (McTainsh and Boughton, 1993). This, combined with very low rates of soil formation (section 8.4.5), means that the Australian environment's tolerance for soil erosion is extremely low.

1.3 Long-term records of environmental change

With soil being such a precious resource in Australia, a detailed understanding of erosion under past and present environmental conditions is essential if we are to understand the consequences of changes in land use and climate on soil erosion. A comprehension of environmental conditions prior to significant human impact is essential if we are to distinguish natural impacts from those resulting from human activity. Knowledge of pre-impact environments provides a much-needed control against which the effects of subsequent human activity may be assessed. In a world in which the biophysical environment has been transformed to a large extent by human activity, perhaps the only means of obtaining reliable experimental control is to examine past conditions. But how much 'control' does examining the pre-European period in Australia provide? Despite the arguments of some that little environmental change took place in the millennia before European settlement in Australia, there is a growing body of evidence that significant climatic fluctuations occurred in the last two millennia. The documented cycles of flood and drought that characterised central eastern Australia during the historical period, for example (Warner, 1987), surely did not begin with the arrival of Europeans? What impact did these processes have on the Australian environment prior to European settlement?

Unfortunately, information on sub-millennial scale environmental variability and long-term human impacts cannot be provided by the instrumental data or historical records at our disposal. The alternative is to turn to proxy records of past environments. The adoption of this approach has led to increasing efforts worldwide to apply palaeoenvironmental techniques to address environmental concerns. This research focus joins a range of collaborative international initiatives, including the International Geosphere Biosphere

Program project BIOME 300, whose aims are to reconstruct changes in global land cover over the past 300 years.

In this study, a *c*. 2000 year palaeoenvironmental record has been developed from Tocal Homestead Lagoon in the lower Hunter valley of central eastern Australia. Detailed investigation of the sedimentary infill of the lake has provided the highest resolution record of late Holocene environmental change yet to be obtained from Australia. In particular, the reconstruction of sediment flux to the basin for the last *c.* 2000 years provides perhaps the best opportunity yet to examine the impact of European settlement on the environment in the early 19th century. There is surprisingly little consensus on the nature of the environmental change that followed European settlement in Australia. Recent work by Butzer and Helgren (2005) joins arguments developed over the last two decades by some Australian researchers that propose that the environmental changes following European settlement were relative minor and that, in many cases, significant impacts were not experienced until as late as the 20th century. As discussed later in this thesis, this view may be coloured to some extent by the predominance of pollen-based palaeoenvironmental records, which appear to indicate only minor shifts in vegetation assemblages at this time. Much of this research has also been plagued by problems in dating environmental change from the period of European settlement. In contrast, recent well-dated records of sedimentation from Australia have suggested that the most dramatic environmental changes were coincident with the arrival of Europeans in the landscape. In this study, major innovations in dating recent sediments (using a wide range of independent techniques) have enabled the horizon of European settlement in the Tocal record to be accurately and precisely determined and, as a result, rates of environmental change from that time onwards may be interpreted with confidence.

The Tocal catchment has been farmed near-continuously since *c*. 1820, with the site used for crops, viticulture, cattle ranching, sheep, dairying, education and tourism at various stages in its history. As the farm has been occupied by only three families since European settlement, good records of the history of land use have been preserved. This site has, therefore, provided a rare opportunity to study the impacts of range of human activities on catchment degradation at a single location. Furthermore, a range of archaeological evidence suggests that the catchment of Tocal Homestead Lagoon was a place of Aboriginal occupation in prehistory. It is possible then that some imprint of Aboriginal activity at Tocal may be preserved in the lake sediment record. With the Tocal sedimentary record extending back to *c*. 120 BC, it is possible that a continuous record of human activity under separate and contrasting cultures may be produced from the site.

1.4 Research aims

The primary aim of this research is to develop a high-resolution record of catchment sediment yield from Tocal Homestead Lagoon for the last *c.* 2000 years. In doing so, answers to the following questions may be provided:

1) What was the effect of European settlement on soil erosion, and how large was this impact in comparison to rates of erosion throughout the late Holocene?

2) How have rates of soil erosion varied during the last *c.* 175 years of European land use? What has been the impact of a range of land uses on the catchment erosion?

3) What other factors have influenced soil loss? Specifically, what has been the impact of cycles of floods and droughts, and fires? Were climatic variations before European settlement recorded at Tocal and did they influence catchment denudation?

4) Does the lake sediment record preserve any mark of pre-European human activity? Can information on the chronology and nature of Aboriginal occupation at the site be obtained from the deposits?

These questions will be addressed by:

1) Constructing a continuous, well-calibrated and high-resolution chronology for the late Holocene at Tocal using AMS ^{14}C, ^{210}Pb, ^{137}Cs, palaeomagnetic and pollution stratigraphy dating methods.

2) The analysis of the physical and geochemical properties of the Tocal lake sediments, in order to obtain proxies of environmental shifts through the late Holocene.

3) The reconstruction of sediment flux to the lake basin during the last *c.* 2000 years, and the calculation of a detailed long-term record of catchment-wide soil erosion.

4) The quarrying of archival sources for information on past land use at the site and in the immediate environment, so that the timing and impact of different land uses may be understood.

1.5 Thesis outline

Chapter 2 presents a description of the biophysical environment of Tocal Homestead Lagoon, its catchment, climate and regional setting. A review of Holocene-scale environmental change in eastern Australia is provided in Chapter 3, with a particular focus on the late Holocene period in central eastern Australia. Our current knowledge of the impact of European settlement on New World environments is also presented, with discussion of the main debates which characterise this field. Chapter 4 presents the history of land use at Tocal and its immediate environment, with a discussion of the archaeological record of Aboriginal occupation in the Hunter valley region and a detailed chronology of European land use at the site since *c.* 1820. Chapter 5 details the methods used in this study. In Chapter 6, the results of the field and laboratory work undertaken in this project are presented. Chapter 7 discusses the results of the dating experiments in the project, which are then used to devise a *c.* 2000 year chronology for the Tocal environmental record. The high-resolution lake sediment record of environmental change at Tocal for the last c. 2000 years appears in Chapter 8. This chapter brings together the findings from this research to address the questions raised section 1.4. Chapter 9 presents the conclusions of this research, and identifies further research opportunities.

Chapter Two

The Environment of Tocal

2.1 Regional geology and geomorphology

The Hunter valley of eastern New South Wales extends almost 200 km inland from the east coast of Australia, encompassing an area of 22 000 km^2 (Figure 2.1B). The geological structure of the valley is very complex. The dominant feature is the northwest-trending Hunter–Mooki Thrust; a zone of thrust faults that dates from the Late Permian or Early Triassic. This separates the two major geologies in the valley. Heavily deformed, folded and faulted Carboniferous rocks outcrop in the northeast, while folded Permian rocks crop out to the southwest (Engel, 1966). The Tertiary in the Hunter is generally considered to be a period of long-term erosion and episodic volcanism (Drysdale *et al.*, 2000), during which time the Hunter River entrenched a course to the coast. Since the Tertiary, the Hunter River has become alluviated, burying its ancient bedrock channel (Drysdale *et al.*, 2000).

By around 20 000 BP, the Hunter River had excavated a valley approximately 50 m deep and 5 km wide (Boyd *et al.*, 2002). Following the most recent sea level rise, which was completed by *c.* 6000 cal. BP, the lower Hunter valley was infilled with estuarine deposits, which extend into the lower portions of the Williams and Paterson valleys, left-bank tributaries of the lower Hunter (Drysdale *et al.*, 2000). Upstream of the infilled estuary, the late Holocene Hunter River was characterised by a flashy discharge regime which transported gravel through a heavily forested floodplain (Boyd *et al.*, 2002).

The highest elevations in the Hunter valley are found to the northwest in the Mount Royal Range, where the Barrington Tops plateau, a massif of Permian granite, lies between 1000 and 1500 m above sea level. The major rivers of the valley, including the Paterson, form a radial drainage pattern centred on Barrington Tops (Nashar, 1964). Pain (1983) has interpreted this as superimposed drainage on an ancient volcano, whose centre was probably close to Mount Barrington.

2.2 Tocal Homestead Lagoon

The low relief floodplains of the rivers of the Hunter valley possess numerous lakes. The Paterson and lower Hunter River valleys alone contain 42 such water bodies (Timms, 1987), one of which is Tocal Homestead Lagoon. Tocal Homestead Lagoon is a relatively small (6.8 ha), permanent water body situated 10 km north of the city of Maitland in the valley of the Paterson River, a left-bank tributary of the Hunter (Figure 2.1B). The lake lies adjacent to Webber's Creek near its confluence with the Paterson River at an elevation of approximately 22 m above sea level (32.6247°S, 151.5919°E). Modern observations of the lake suggest it to be seasonally eutrophic. Blue-green algal (Cyanobacterial) blooms and water hyacinth (*Eichhornia crassipies*) were observed during field work in the summer of 1998, the latter having been present in problematic concentrations in the lake in previous years (Archer, A.C., Principal of Tocal Agricultural College, personal communication, 2003).

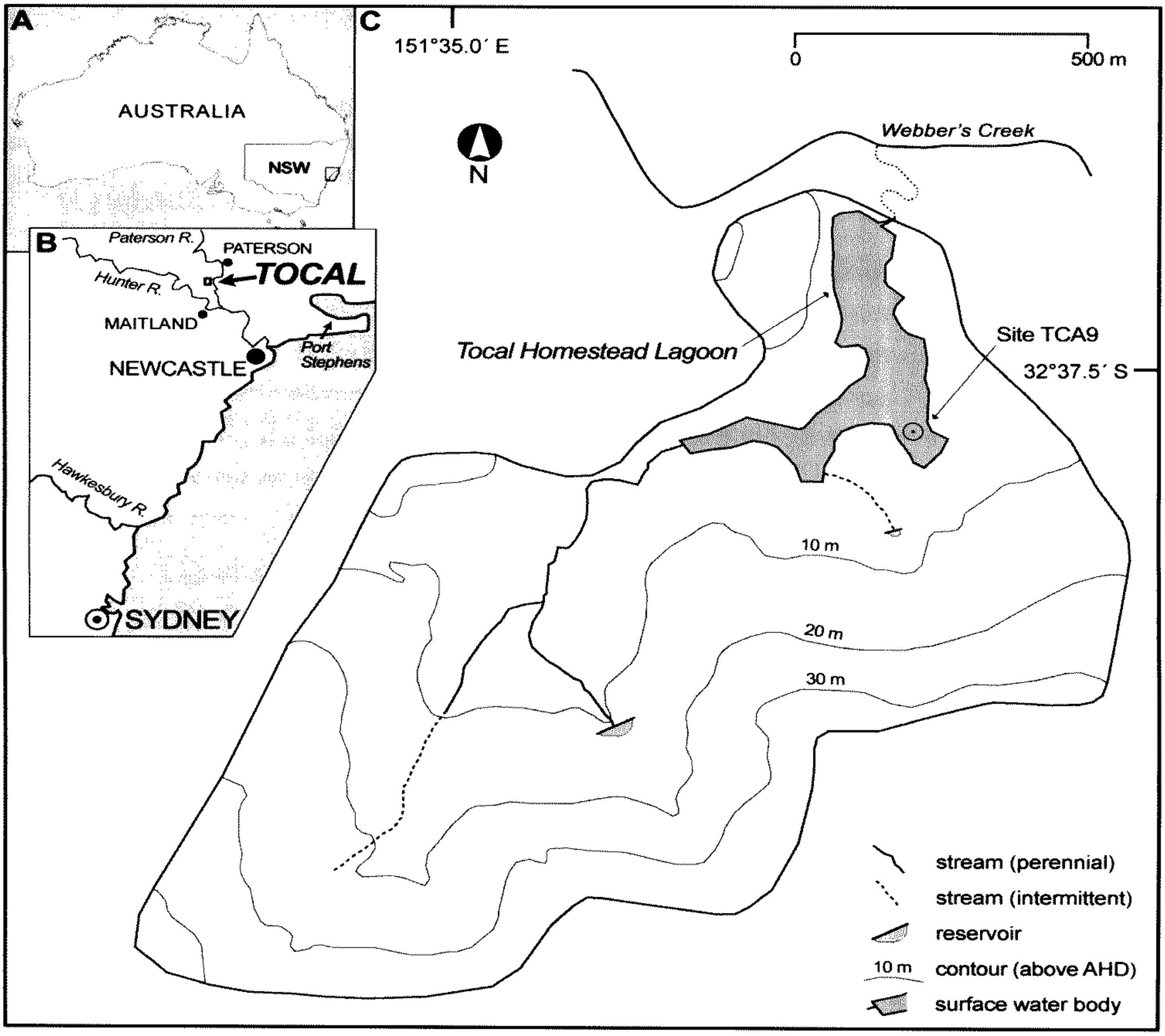


Figure 2.1 The catchment of Tocal Homestead Lagoon and its situation in New South Wales (NSW), eastern Australia, showing the locations discussed in the text. Mapping based on field surveys, 1993 aerial photographs and the unpublished Geographical Information System established for Tocal by Agriculture New South Wales. From Cook *et al.* (in press).

The lake is a 'lateral lake' (Hutchinson, 1957) or a 'blocked valley lake' (Timms, 1987), having formed following the rapid aggradation of a river levee and subsequent damming of a tributary. This probably occurred in the mid Holocene as sea levels approached their present heights (Thom and Roy, 1983) and as streams aggraded to form their modern floodplains. Support for this belief comes from the radiocarbon age of 6840±50 BP obtained from a depth of 3.18–3.16 m in Tocal Homestead Lagoon and from a radiocarbon age of 5277±54 BP, the oldest so far obtained from blocked valley lakes in the Hawkesbury catchment, less than 100 km to the south (Gale, 2002). The lake retains its dendritic plan form, with the main basin aligned approximately north–south along the former channel (Figures 2.1C, 2.2 and 2.3), and a much smaller and shallower limb to the southwest of the main basin (Figures 2.4 and 2.5). Although water depth in the lake varies throughout the year, during sampling of the lake sediments in 1998 the average depth was 0.80 m, with a maximum of 1.38 m (Figure 2.6). During periods of high rainfall, water depths in the lake have exceeded 3 m.

Tocal Homestead Lagoon has a small, modern artificial outlet at its northern end, through which water drains into Webber's Creek under flood conditions (Figures 2.1C and 2.7). An examination of aerial photographs suggests that the outlet was constructed in the 1960s, probably to maintain vehicular access to Tocal Homestead from Tocal Road (Figure 2.3). Floods overtopping the levee dam prior to the 1960s are thought to have been rare, and the lake therefore has a high trap-efficiency under natural (pre-artificial outlet) conditions. Since its formation, the lake has effectively acted as an enclosed sedimentary basin, trapping material eroded and transported from the catchment.

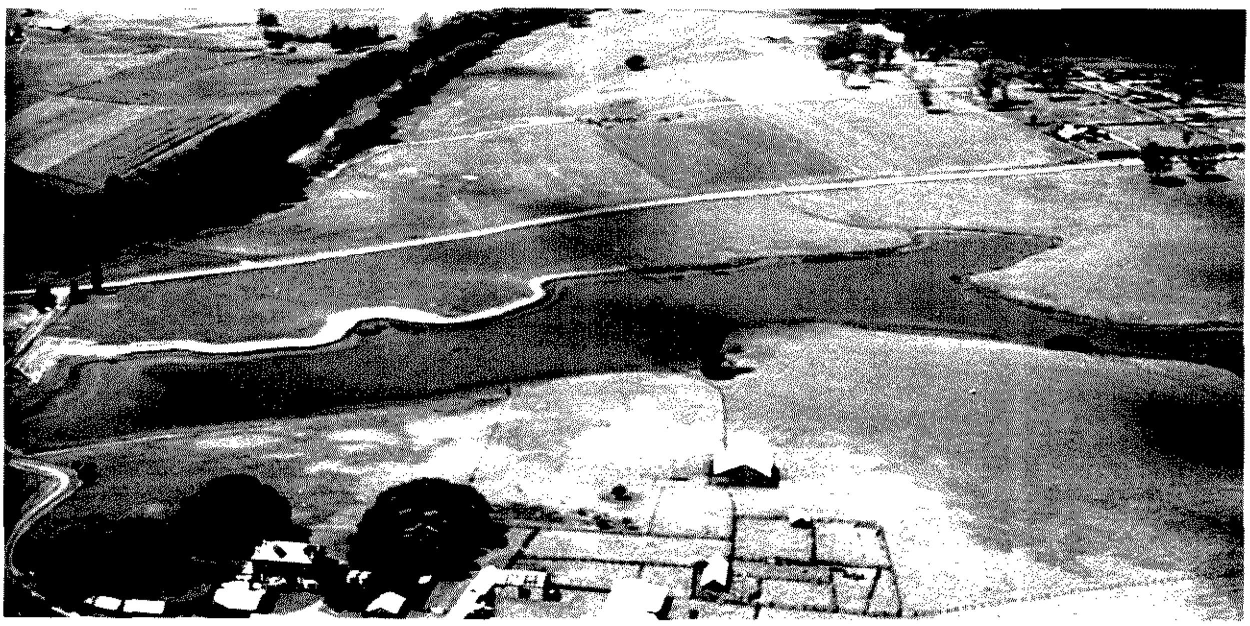


Figure 2.2 Tocal Homestead Lagoon, Paterson, eastern Australia. The Tocal Homestead and associated buildings may be seen in the foreground, while the nearby Paterson River is seen in the background, date unknown (Source: Tocal Library Archives).

Figure 2.3 The 1996 aerial photograph of Tocal, showing the boundary of the lake's catchment (dashed line), with the college buildings in the southeast corner and Tocal Homestead and other early colonial structures in the northwest corner. North lies at the top of this image.

Figure 2.4 Tocal Homestead and associated buildings in the foreground, the southwest limb of the lagoon and the C.B. Alexander College campus in the background, date unknown (Source: Tocal Library Archives).

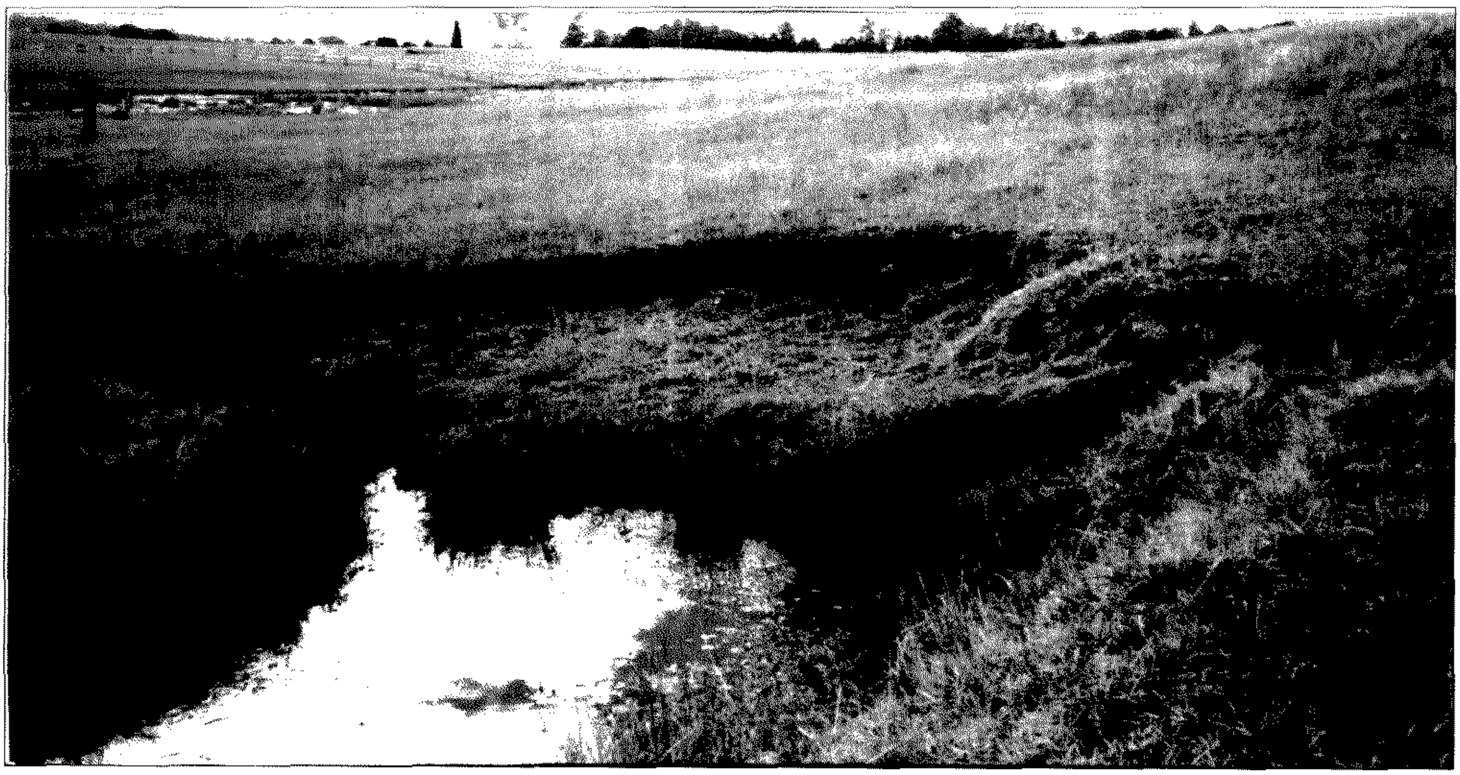


Figure 2.5 The unnamed creek that drains the southwest of the catchment into Tocal Homestead Lagoon, with some slumping of the outer bank, 1998. The direction of the drainage is into the page.

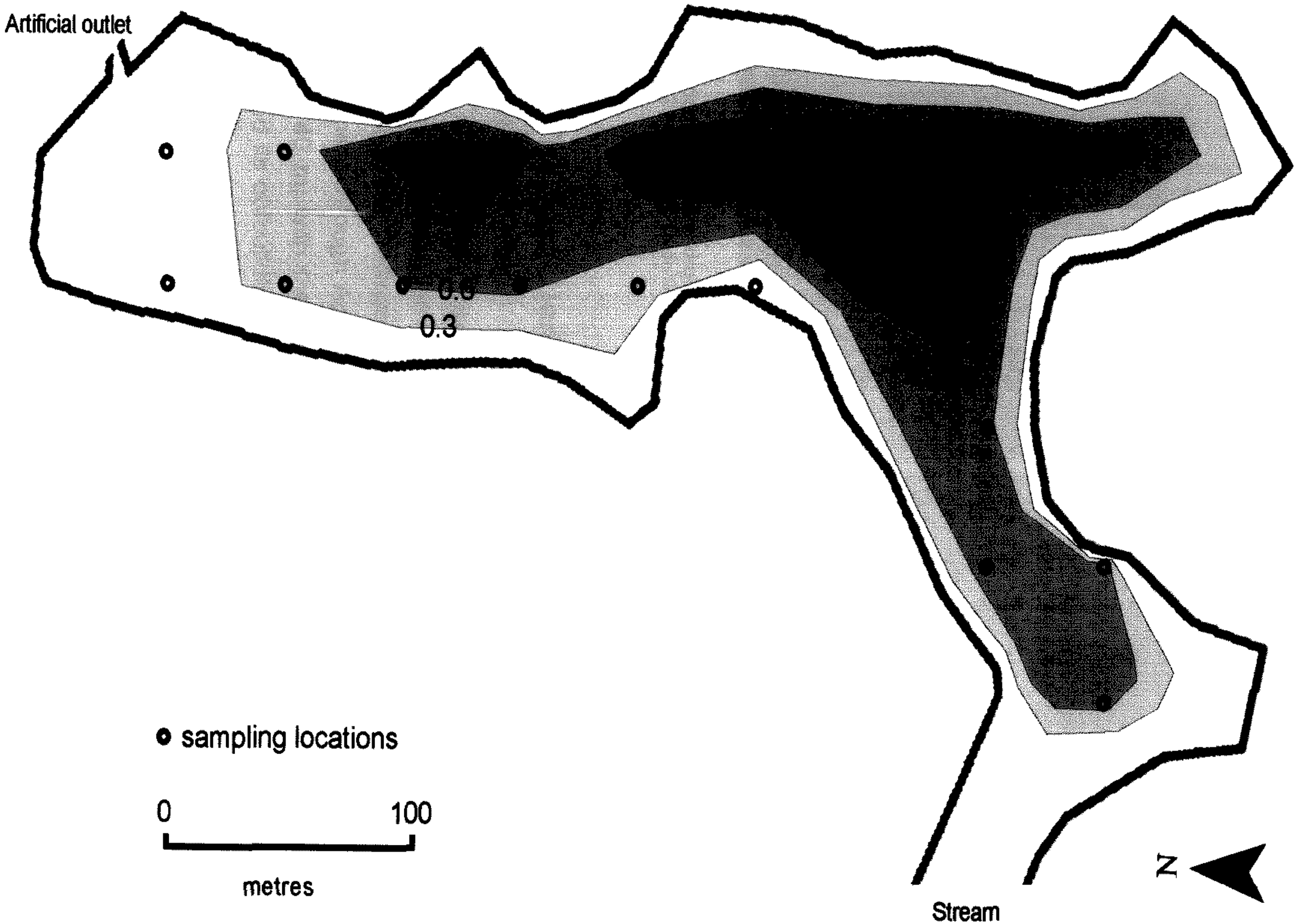


Figure 2.6 Bathymetry of Tocal Homestead Lagoon, based on measurements of water depths recorded during field work in 1998 (units in metres). Measurements made at sites after this date have been corrected to the water depths from the 1998 field season.

Figure 2.7 Modern flood-level drainage outlet of the northern end of Tocal Homestead Lagoon via a concrete culvert, running beneath the main access road to the Tocal Homestead, 1998.

2.3 The Tocal Homestead Lagoon catchment

2.3.1 Catchment properties

The Tocal Homestead Lagoon catchment forms part of the 2200 ha property of the C.B. Alexander Agricultural College (Figure 2.2), which is administered by the New South Wales government through Agriculture New South Wales. The catchment has a relief of approximately 40 m, with the highest elevations being the small hill upon which Tocal Homestead and its outbuildings are located, the crest to the south of the lake where the college campus was constructed, and a small segment of woodland in the far southwest corner of the catchment. The total catchment of the lake consists of two sub-catchments totalling 93 ha, one immediately surrounding the lake, and a much larger area to the southwest of the lake (Figures 2.1 and 2.2).

The sub-catchment immediately surrounding the lake drains directly into the basin along relatively short and steep slopes, and via two ephemeral drainage lines that enter the lake on its southern shore. The larger southwest sub-catchment delivers water and sediment via a shallow intermittent stream network into the southwest limb of the lake.

Foster *et al.* (1990) and Dearing and Foster (1993) have shown that sediment delivery is most efficient in catchments that are small relative to their lake area. In a little catchment such as that of Tocal, there are direct linkages from slopes to the lake basin. Sediment storage on these slopes is negligible, and eroded sediment is transported efficiently into the basin. The lake is sedimentologically isolated from the nearby Paterson River (section 8.2.3). The site is therefore well suited to the study of catchment soil erosion and land use history, with the lake sediment record likely to contain a continuous and undisturbed record of catchment sediment yield. As such, the sediments within the lake are expected to represent a direct and near-instantaneous response to catchment events.

The North Coast railway line from Maitland to Dungog (constructed in 1901) bisects the catchment, separating approximately 17 ha of the southwestern catchment from the remainder (Figure 2.3). This area drains into the eastern part of the catchment via small concrete culverts located beneath the railway line. The eastern edge of the catchment (17 ha), on the far side of Tocal Road, forms part of a separate agricultural property owned by the college called Glendarra (Figure 2.2). Although runoff from this area drains into the lake, sediment transport may be significantly interrupted by the road and its associated drainage works.

2.3.2 Geology

The Tocal catchment is composed of Permian rocks of the Branxton Formation, part of the basal unit of the Permian Maitland Group (Rose *et al.*, 1966). The Branxton Formation consists of sandstones, sandy siltstones and conglomerates, and is characterised by large quantities of marine fossils, particularly shells, polyzoa and crinoids (Nashar, 1964). These beds are overlain by a westward thinning sequence of Quaternary alluvium. The lake and its immediate catchment are formed on the alluvium, while the majority of the catchment is underlain by sandstone, interbedded with tillite and, to a lesser extent, fossiliferous conglomerate. The tillite consists largely of poorly sorted, angular boulder-sized granite erratics.

2.3.3 Soil

Information on the soils of the Tocal Homestead Lagoon catchment comes from a combination of field observations, soil sampling and analysis of selected profiles, soil unit mapping undertaken by Agriculture New South Wales as part of Tocal's Geographic Information System (GIS) (Figure 3.8 and Table 3.1), and limited sampling and analysis undertaken by the New South Wales Department of Land and Water Conservation (DLWC) in 1994.

Some relationship can be seen between catchment soil types and their underlying parent material, topography and, to a lesser extent, catchment hydrology. Soils in the upslope area in the southwestern portion of the catchment are relatively shallow (<0.50 m deep), gravelly sandy loams with truncated A-horizons (tenosols, lithosols) and little or no soil fauna, which have formed on the predominantly sandstone bedrock. Near drainage lines or on mid slopes, deeper brown earthy sand clay loams have developed (kurosols, duplex podsolic soils). The low relief areas immediately surrounding the lake consist of deeper (>3 m) alluvial soils (rudosols and hydrosols) formed on Quaternary alluvium. Away from the lake, deeper kurosols have formed on the same parent material. Thin lenses of alluvium overlie other low-lying areas towards Webber's Creek. The moderately steep slopes to the south, east and west of the lake consist of chromosols, kurosols and some sodosols, with evidence of gilgai microrelief in some locations. The chromosols are found on the mid-slopes further away from the lake, formed on sandstone parent material. As with the shallow rudosols, these soils are characterised by truncated A-horizons, suggesting either significant erosion, soil compaction, or both.

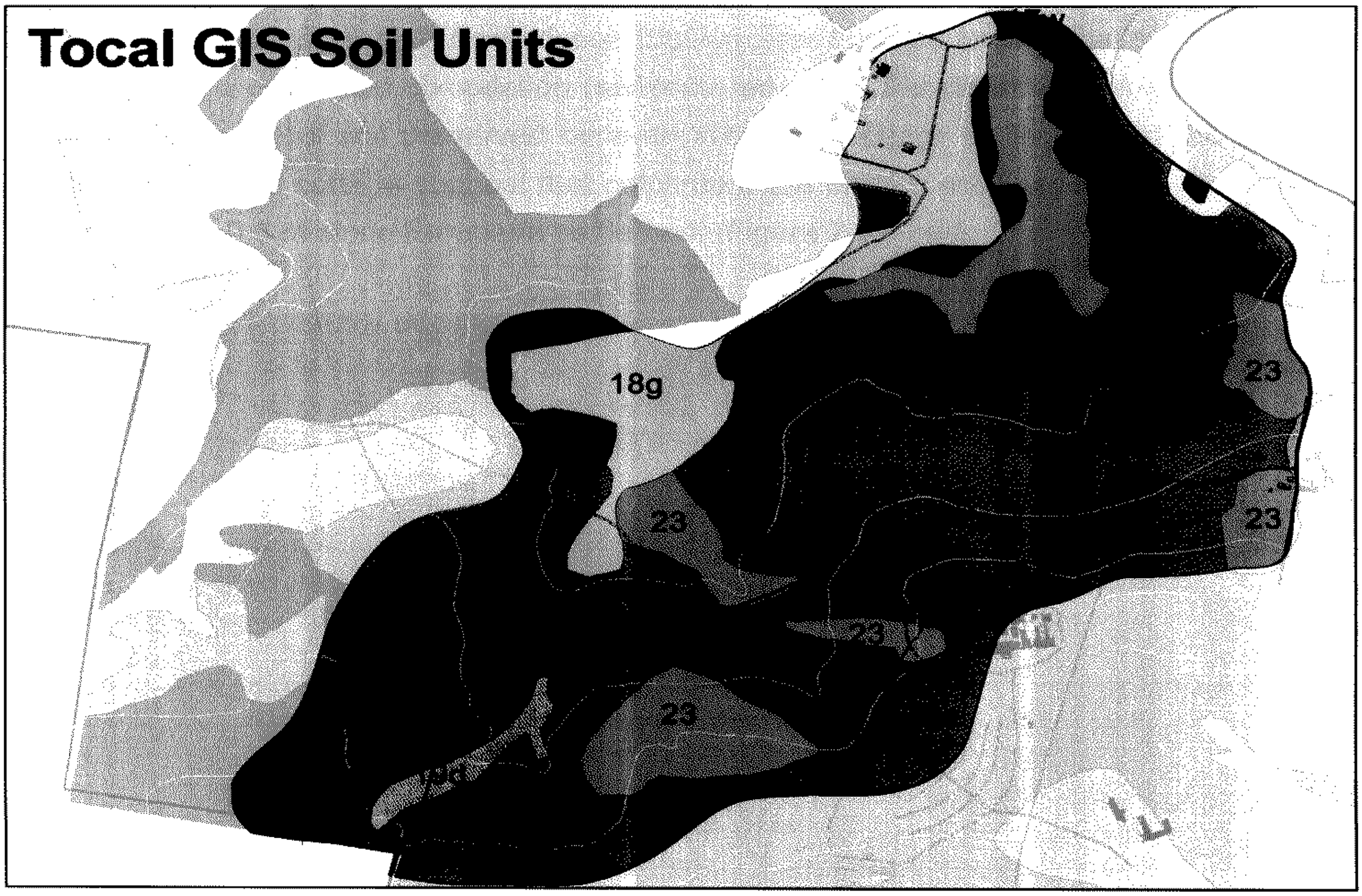


Figure 2.8 Soil units identified in the Tocal Homestead Lagoon catchment as part of the Agriculture New South Wales GIS (used with permission). See Table 2.1 for key.

Table 2.1 Tocal soil mapping units, from the Agriculture New South Wales GIS (used with permission).

Soil unit	Description	
	Northcote (1979)	**Australian Soil Classification (Isbell, 2002)**
4w	Uf 6.32	Hydrosols, rudosols
17w	Uf 6.32	Hydrosols
18	Dy Dr 4.1+2 Db Dd 3.1+2	Kurosols, rudosols
l8g	Dy Dr 4.1+2 Db Dd 3.1+2	Chromosols, some vertisols, some gilgai micro-relief
19d	I 9d, Dy Dr 4.3+4 Db Dd 3.3+4	Kurosols, periodically inundated
20	Dy Dr 4.5+6 Dd Db 3.5+6	Kurosols
22	Dy Dr 5.1+2 Dd Db 4.1+2	Kurosols
23	Dy Dr 5.3+4 Dd Db 4.3+4	Kurosols, occasional sodosols
25	Dy 5.7+8	Kurosols, chromosols

A soil sequence sampled by DLWC in 1994 as part of a regional mapping program to identify areas of acid-sulphate soils, reported no 'Actual acidity' or 'Potential acidity', but classified the soil as sodic. Dispersible soils are relatively common in the Hunter valley region. The sodicity–dispersibility relationship in such soils often contributes to increased soil erodibility (Ford *et al.*, 1993). Preliminary soil chemical analysis undertaken by Cook (1998) demonstrated the sodicity of some of the soils in the catchment to be in the range of 'dispersible soils', as defined by Ford *et al.* (1993).

Field observations of thin, truncated soil profiles and bedrock exposures suggest that extensive soil erosion and compaction has occurred within the catchment. A comparison between the depth of soil preserved beneath Webber's cottage at Tocal, and the same soil adjacent to the cottage suggest that catchment soil profiles may have been truncated by as much as 0.5 m since European settlement (Figure 2.9).

Figure 2.9 Webber's cottage in 1999, Tocal Homestead Lagoon, Paterson, eastern Australia. The green line marks the approximate location of the top of the soil profile preserved beneath the building's floor boards.

A kurosol sampled from near the campus buildings of Tocal by Agriculture New South Wales has a truncated A-horizon, with only *c.* 0.30 m of topsoil (Figure 2.10). Low relief areas adjacent to the lake shore and creek are characterised by significantly deeper soils, which may include some contribution by material transported from further upslope. Interestingly, no geomorphological evidence of soil erosion, such as gullying or erosional scars, is visible in the catchment today.

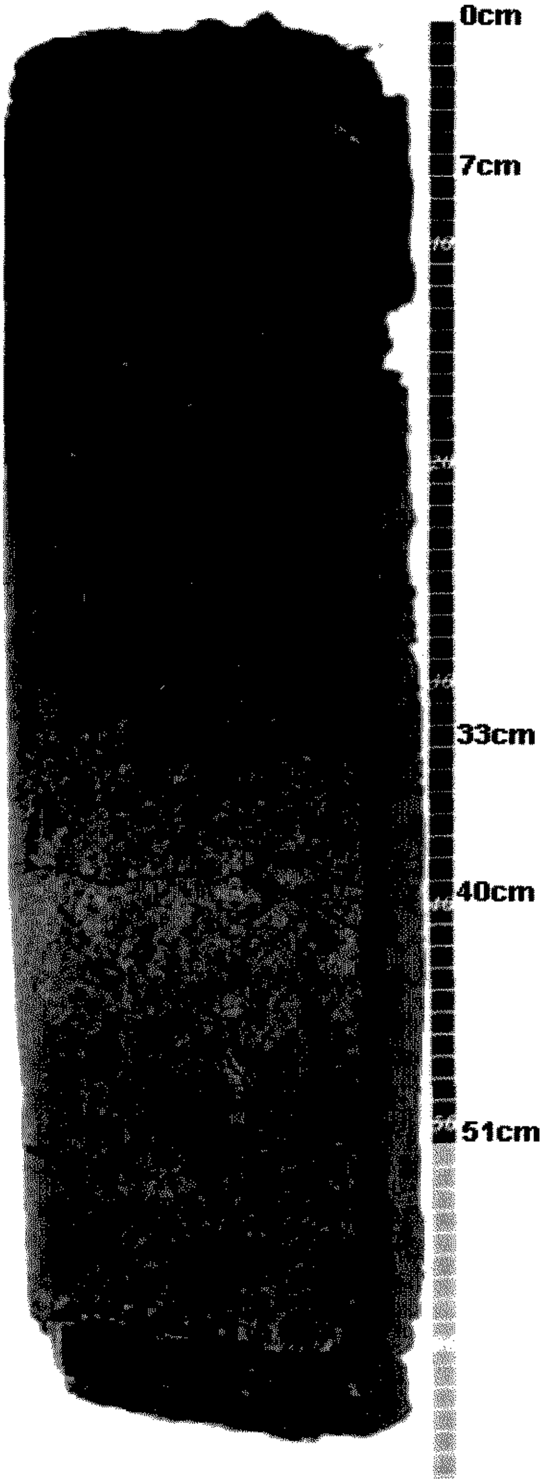


Figure 2.10 Kurosol soil profile from the catchment of Tocal Homestead Lagoon, Paterson, eastern Australia. Only *c.* 0.30 m of loamy topsoil remains at this location today. Image provided by Tocal Agricultural College.

2.4 Vegetation

Early accounts by the first European settlers of the Paterson River area suggest that the vegetation of the floodplain was forest dominated by red cedar (*Toona ciliata*) and, to a lesser extent, kurrajong (*Brachychiton populneus*) (see Chapter 4). What little native forest that exists in the region today consists mainly of spotted gum (*Corymbia maculata*), grey box (*Eucalyptus moluccana*) and forest red gum (*Eucalyptus tereticornis*). The catchment of Tocal Homestead Lagoon contains little woodland today, with discrete clusters of native species being concentrated in the southwest of the catchment and along the catchment divide to the south, with a patch of remnant forest located immediately north of the lake on the southern edge of Webber's Creek. The pockets of open forest include spotted gum (*Corymbia maculata*), grey ironbark (*Eucalyptus paniculata*), grey gum (*Eucalyptus punctata*), forest red gum (*Eucalyptus tereticornis*) and forest oak (*Casuarina torulosa*). Remnant riparian rainforest vegetation on the banks of Webber's Creek and on the eastern banks of the Paterson River consists of Moreton Bay fig (*Ficus macrophylla*), flooded gum (*Eucalyptus grandis)*, red cedar (*Toona ciliata)*, white cedar (*Melia azedarach* var. *australasica*), weeping lillypilly (*Waterhousia floribunda*) and broad-leaved paperbark (*Melaleuca quinquenervia*) (Laffan, 2003). Beyond the immediate riparian zone, river oak (*Casuarina cunninghamiana*) and water gum (*Tristaniopsis laurina*) are found today.

The vast majority of the catchment surface is vegetated solely by various native and introduced grasses, an exception being the gardens and immediate area of the Tocal Homestead which contains numerous exotic trees such as willow (*Salix babylonica*), poplar (*Populus fremontii*) and garden plants.

2.5 Climate

2.5.1 Introduction

A useful discussion of the climate at Tocal has been presented by Archer and Muddle (1988). Three Bureau of Meteorology weather stations operate in the area. One station is located within the catchment, approximately 500 m from the lake, and began operation on 19 November 1967 (Table 2.2). The other two are at Paterson Post Office (3 km north of the site) and Maitland West (approximately 10 km south of Tocal). The Maitland West site has the benefit of an unusually long record that dates from 1853 (rainfall records began here in 1858). Unfortunately, mating the three records is not straight forward. Climate varies considerably throughout the Hunter valley, and localised effects are often quite important. Such local-scale variations may be significant in influencing the climate at

Tocal. For example, a comparison of the mean daily rainfall record at Tocal with that from the Paterson Post Office station (Table 2.3) shows a similar yearly trend in the distribution of rainfall, but with the Paterson Post Office generally recording more rainfall each month, particularly through summer.

2.5.2 Temperature

Tocal has a Cfb climate based on the Köppen classification system, characterised by a warm temperate climate with hot summers and no discernible dry season. The site's proximity to the coast has the effect of reducing summer maximum temperatures compared to other locations in the Hunter valley (Archer and Muddle, 1998). Summer months are warm to hot; January is the warmest month with a mean daily maximum temperature of 29.4°C and a mean daily minimum of 17.5°C. Winter months are cool and the site experiences an average of ten frosts per year (Archer and Muddle, 1988). The coldest month on average is July, with a mean daily maximum temperature of 17.1°C and a mean daily minimum of 6.1°C.

2.5.3 Wind

Regional wind patterns are controlled by the Southern Maritime and Tropical air streams, with the additional influences of the Tropical Continental air stream in the winter months and the Southern Continental air stream from spring into summer (Archer and Muddle, 1988). In some winters the region experiences Polar Maritime air streams; these have resulted in snow falls on the Mount Royal Range north of Tocal (Archer and Muddle, 1988). Katabatical processes are also important in the region's wind climate, and are responsible for strong westerly winds in the summer months along the bottom of the middle and lower Hunter valley.

2.5.4 Rainfall

Most of the rainfall in the region occurs during the summer and autumn (Table 2.2), with subtropical and coastal tropical low-pressure systems being largely responsible. Winter rainfall is typically generated by low-pressure systems and cold fronts moving across the continent from the Tasman Sea (Anon., 1985). Rainfall in the area around Tocal is on average higher than in other areas of the Hunter valley due to localised orographic effects. These effects are even greater further upvalley, with Paterson Post Office typically recording higher rainfall than Tocal over the period 1968–2001 (Table 2.2).

The mean annual rainfall at Tocal is 925.8 mm, with the wettest months being January, February and March when rain occurs in association with onshore winds. July, August and September record the lowest rainfalls on average, with rainfall approximately half that of the wettest months. Data for rainfall erosivity in the region show a distinct tendency for higher erosivity in summer months when compared to the winter months, with 71% of the annual rainfall erosivity calculated for nearby Chichester Dam (approximately 15 km northeast of Tocal) occurring in the summer (Rosewell and Turner, 1992). This is possibly due to the occurrence of higher intensity storms in the summer months.

Table 2.2 Climatic statistics for Bureau of Meteorology station 061250 Paterson, New South Wales (Tocal) (Elevation 30 m AHD, Latitude: 32.63°S, Longitude: 151.59°E) and rainfall data from station 061096 Paterson Post Office, New South Wales (Elevation: 21 m AHD, Latitude: 32.60°S, Longitude: 11.62°E).

Month	Jan	Feb	March	April	May	June	July	Aug	Sept	Oct	Nov	Dec
Paterson, New South Wales (Tocal), 1968–2001												
Mean daily maximum temperature (°C)	29.4	28.6	26.8	24.2	20.6	17.6	17.1	19.0	22.0	24.6	26.2	28.9
Mean daily minimum temperature (°C)	17.5	17.4	15.6	12.5	9.9	7.5	6.1	6.6	8.7	11.3	13.7	16.1
Mean daily rainfall (mm)	3.9	3.9	3.9	2.6	2.6	2.2	1.4	1.1	1.6	2.2	2.7	2.6
Mean daily evaporation rate (mm)	6.1	5.2	4.2	3.4	2.5	2.3	2.6	3.6	4.5	5.4	6.3	7.2
Paterson Post Office, New South Wales, 1901–2001												
Mean daily rainfall (mm)	3.6	3.9	3.8	3.3	2.4	3.1	2.3	1.9	2.1	2.3	2.2	3.4

Table 2.3 A comparison of the mean daily rainfall data recorded at Bureau of Meteorology station 061250 Paterson, New South Wales (Tocal) with that from station 061096 Paterson Post Office, New South Wales for the period when they were both in operation, 1968–2001.

Month	Jan	Feb	March	April	May	June	July	Aug	Sept	Oct	Nov	Dec
Mean daily rainfall (mm), Paterson, New South Wales (Tocal), 1968–2001												
	3.9	3.9	3.9	2.6	2.6	2.2	1.4	1.1	1.6	2.2	2.7	2.6
Mean daily rainfall (mm), Paterson Post Office, New South Wales, 1968–2001												
	5.2	4.6	5.3	3.3	3.4	3.0	1.6	1.5	1.8	2.7	3.0	3.3

2.5.5 Long-term rainfall trends

By combining the Tocal rainfall dataset with that from Paterson Post Office, a 100 year record may be determined. These records can be combined quite easily; although the Paterson Post Office station on average receives higher rainfall, the annual patterns of rainfall recorded at each station are very similar (Table 2.3). More importantly, the long-term pattern of total annual rainfall at each station should be similar. The annual rainfall at Tocal has varied significantly during the period of record keeping (Figure 2.11). Periods of much wetter conditions prevailed during the 1910s, late 1920s, 1950s, 1970s, and the late 1980s to early 1990s (Figure 2.12). Exceptionally wet years in the Paterson River valley were recorded in 1921, 1949, 1950 and 1963, when the annual rainfall was more than 50% greater than the long-term average of 987 mm per year (Figure 2.12). This record can be extended even further back in time, by adding on the rainfall record from the Maitland West station. However, as detailed in section 2.5.1, the distance of this station from Tocal requires that this early segment of the record be interpreted with care (Figure 2.13). This extended record suggests that much wetter conditions prevailed in the region in the early 1880s, and most of the 1890s, while drier conditions reigned through the late 1880s and the 1900s. Exceptionally wet years were recorded in 1879, and from 1892 into 1893, while drought years in the Maitland area were 1882, 1888 (the driest year of the nearly 150 year record with only 481.2 mm of rain recorded) and 1901.

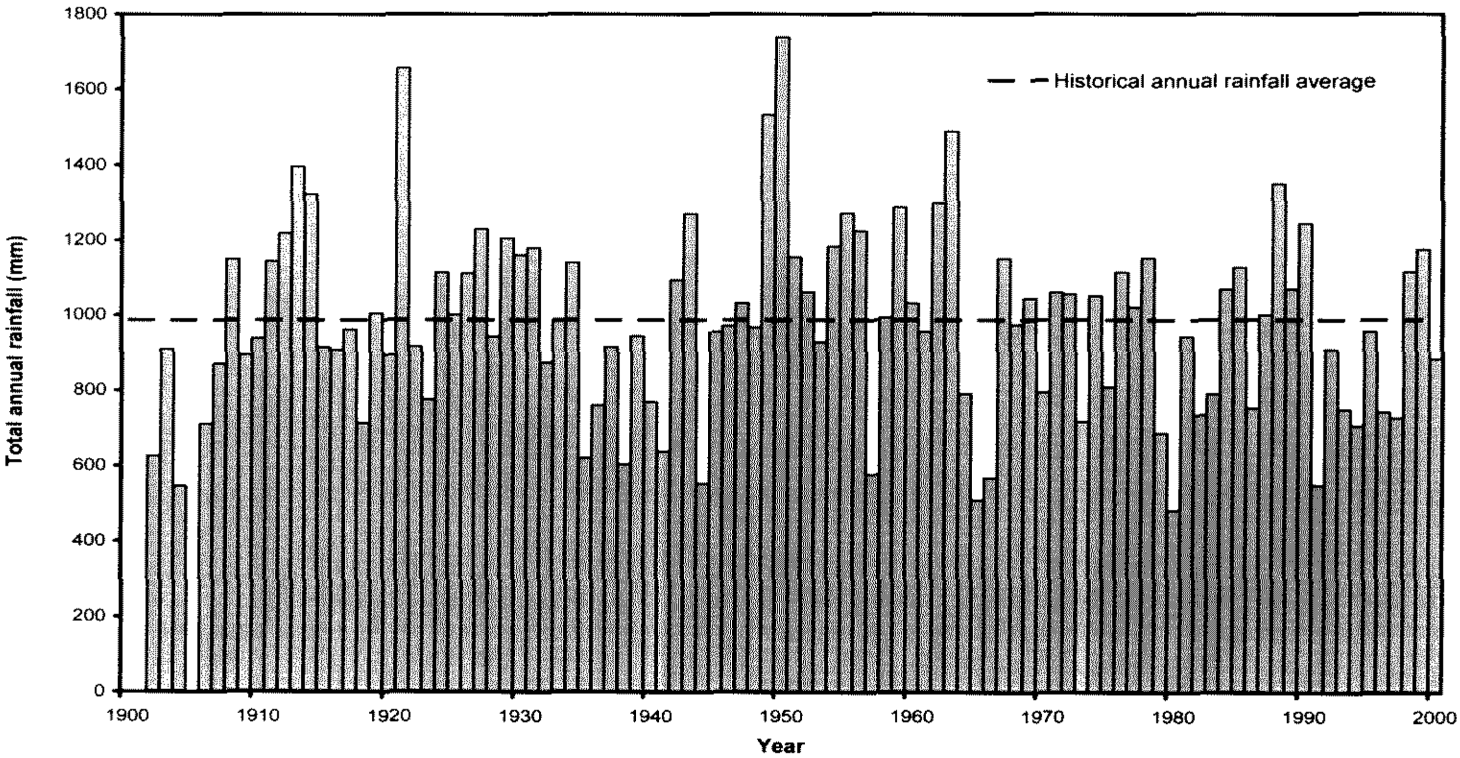


Figure 2.11 Total annual rainfall recorded at the Bureau of Meteorology station 061250 Paterson, New South Wales (Tocal) (1902–1967) and station 061096 Paterson Post Office, New South Wales (1968–2000) and the mean annual rainfall for the period 1906–2000. No annual data are available for the years 1901 and 1905 due to incomplete readings at Paterson Post Office.

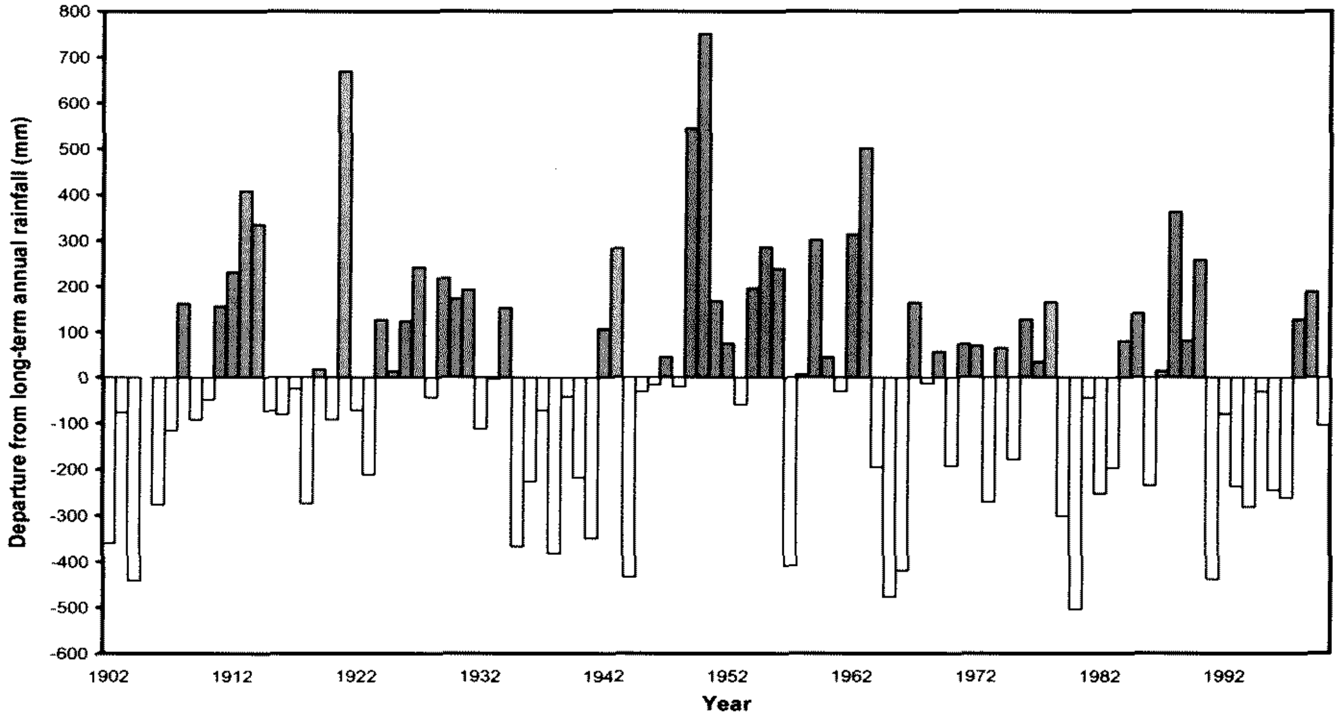


Figure 2.12 Annual rainfall recorded at the Bureau of Meteorology station 061250 Paterson, New South Wales (Tocal) (1902–1967) and station 061096 Paterson Post Office, New South Wales (1968–2000) expressed as the departure from the historical long-term average for the period 1906–2000. No annual data are available for the years 1901 and 1905 due to incomplete readings at Paterson Post Office.

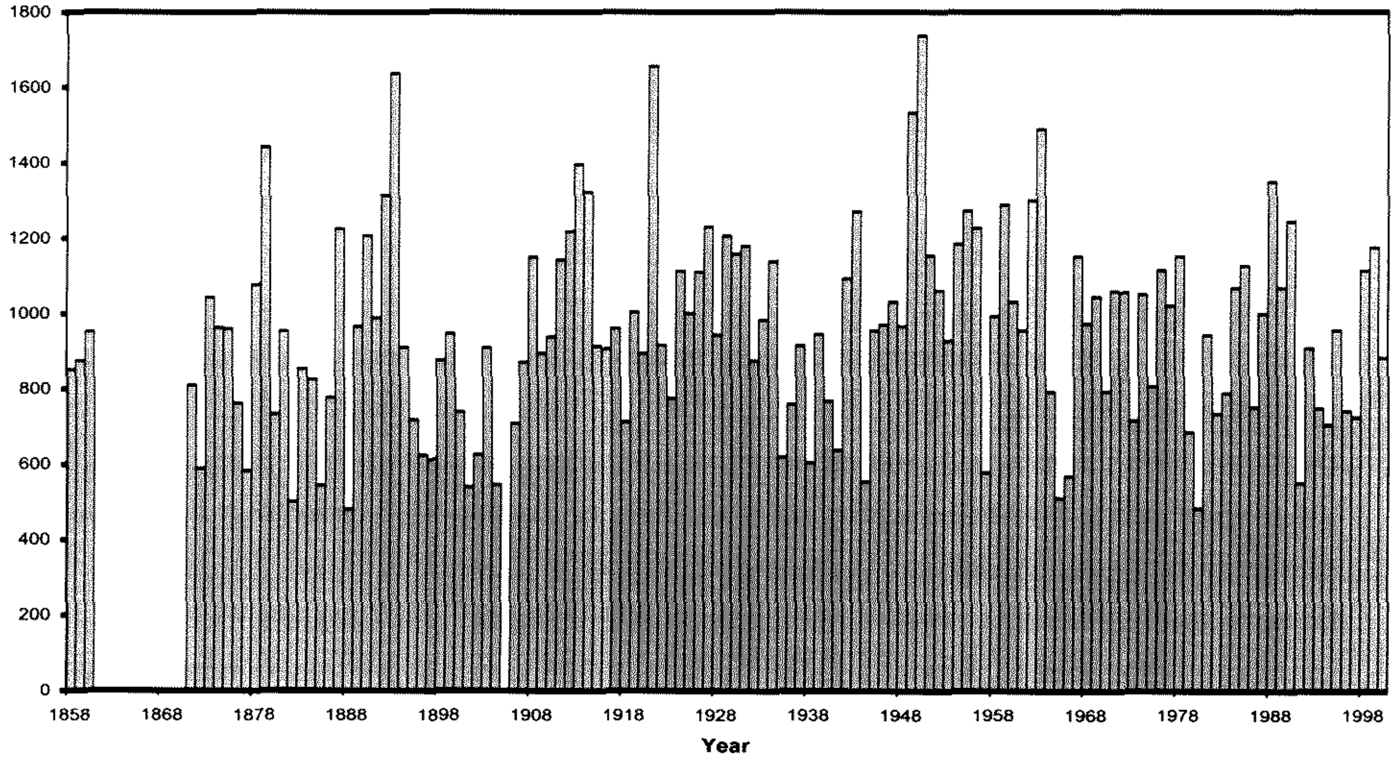


Figure 2.13 Annual rainfall recorded at the Bureau of Meteorology's Maitland West (1858–1902), Paterson, New South Wales (Tocal) (1902–1967) and station 061096 Paterson Post Office, New South Wales (1968–2000). No annual data are available for the years 1861–1870 and 1905 due to incomplete readings at Maitland West and Paterson Post Office.

2.5.6 Droughts

Periods of below average rainfall in the Paterson River valley dominated the late 1890s, the early 1900s, the late 1930s to mid-1940s, the mid-1960s and the early 1980s. Even though the region has a high annual rainfall by Australian standards, droughts have been a problem in the past. The historical record reports major agricultural droughts in the following years: 1798–1799, 1803, 1809–1811, 1813–1815, 1824–1829, 1845–1847, 1895–1899 (for eastern New South Wales from Foley [1957]), 1902–1903, 1918, 1938–1941, 1944, 1957, 1964–1965, 1980, 1984 (in the Hunter valley, from Archer and Muddle [1988]) and 1991. In the 1964–65 drought, the Allyn River stopped flowing at Hilton (approximately 10 km east of Tocal) from 9 January 1965 to July 1965; and the Paterson River stopped flowing upstream at Mount Rivers from 11 December 1964 to 30 January 1965 and from 8 March 1965 to July 1965 (Pattison, 1966). In the 1965 drought, the encroachment of marine waters up the Paterson River as far as Paterson caused a halt to irrigation (Archer and Muddle, 1988). Periods of drier than average years (Figure 2.11) correspond well with the historical records of droughts. However, it appears that agricultural droughts do not always coincide with the calendar year of low rainfall (with agriculturally-defined drought often taking place in the next year).

2.6 Rainfall, runoff and proxy records of floods and droughts

2.6.1 Rainfall and storm losses

By examining the relationship between rainfall, runoff and floods in the region, the capacity of the long-term climatic record to act as a proxy for floods or droughts may be assessed. Runoff processes are highly complex, both spatially and temporally. Numerous models exist that describe the behaviour of runoff from a catchment's surface under given conditions of soil and rainfall (e.g., Horton, 1933; Chorley, 1978; Whipkey and Kirkby, 1978). These models are, however, unhelpful in investigating the relationship between rainfall and runoff over long time spans, in part because of the absence of the necessary instrumental data, but also because of a lack of information on soil, climate and land use conditions. Rainfall records are typically the longest and most complete instrumental records available in eastern Australia. Accordingly, it is simpler, and more appropriate, to develop models of runoff that rely primarily on available rainfall data. Such simplification may be achieved through the use of storm losses to estimate catchment storage.

Although a range of storm loss models is used in Australia (discussed in detail in Cordery [1987]), the most widely used rainfall-based models are the initial loss and continuing loss models of Walsh *et al.* (1991). Walsh *et al.* (1991) have produced estimates of initial storm losses for coastal catchments in New South Wales, using 22 catchments with high-quality flow records located in the nearby Sydney and Shoalhaven regions, as well as using data from the Hunter valley. It is their estimates of initial storm losses that are adopted for this study. By developing a record of initial storm losses for the period of rainfall observations, a proxy record of flood events may be derived. Similar, but less refined approaches have been employed by Page and Carden (1998) and Gale and Haworth (2005) to provide proxy flood records for sites in New South Wales from long-term monthly rainfall totals. In these cases, no account of initial storm losses was made, and runoff is assumed to be a constant fraction of rainfall in each time period.

The storm loss model of Walsh *et al.* (1991) was applied to the long-term record of daily rainfall observations spanning 100 years from the Bureau of Meteorology weather stations at Tocal and Paterson Post Office. A working initial loss value of 50 mm, with error bars of 20 mm, was adopted for calculation in this study. That is, without consideration of antecedent soil conditions, the first 50 mm of rainfall is retained in the soil profile per event. The chosen values are equal to the lowest initial losses for events up to a 10 year Average Recurrence Interval (ARI) and represent a conservative estimate of storm losses. Calculations were also performed on the upper (70 mm) and lower (30 mm) bounds of the initial loss estimates in an attempt to quantify the errors in the assumptions of the model.

2.6.2 Testing the proxy flood record against observed floods

One way in which the initial storm loss proxy record of flood events at Tocal may be evaluated is by comparison with data recorded from observed flood events. We are fortunate to have available a compilation of records of floods at Belmore Bridge gauge, Maitland from 1908 to 1967 (Table 2.4) (Anon., 1968b). It should be noted, however, that the greatest 24 hour rainfalls, and possibly the greatest floods, recorded in the Hunter valley took place before the start of this record, in March 1871 and March 1893 (Geary and Erskine, 1984). The river gauge is on the Hunter River, approximately 10 km from the Tocal site. Although it may have experienced a record of floods similar to the Tocal site, the upper Hunter and Paterson River catchments differ significantly and some differences in their hydrological record may exist. The flood compilation records floods as having occurred at the Belmore Bridge gauge when river stage equals or exceeds 27 feet (8.23 m). Interestingly, these data show that most floods take place during winter, when the average daily rainfall is nearly half that of the summer months (fewer days with rainfall per month), though evaporation rates are greatly reduced. This may reflect the impact of the high magnitude thunderstorms that sweep through during the winter months delivering heavy rainfalls, even though the average winter day has much lower rainfall.

The comparison of the storm loss models and the Belmore Bridge stage record is presented in Figures 2.14–2.16. The number of days per year when notional storm losses were exceeded has been plotted against the height of peak flows at Belmore Bridge above 26 feet (stages of 27 feet or more are defined as floods). Some years recorded more than one flood at Maitland. A way of representing these years was needed that would be directly comparable with the storm loss record, which is of annual resolution only. For years when multiple floods were recorded, the heights of the peak flow above 26 feet for each flood in a year were added together.

Table 2.4 Flood events at Belmore Bridge, Maitland, Hunter River, New South Wales from 1908 to 1967. Modified from Anon. (1968b). Floods are defined as occurring when the stage equals or exceeds 8.23 m.

Flood No.	Year	Month	Maximum height (m)	Peakflow (m^3s^{-1})
1	1908	February	8.99	1671
2	1910	January	9.45	2039
3	1913	May	11.28	4587
4	1920	July	8.69	1444
5	1921	July	9.45	2039
6	1921	July	9.30	1897
7	1926	March	9.30	1911
8	1927	April	8.23	1104
9	1929	September	8.69	1444
10	1930	June	11.35	4276
11	1931	April	9.60	2166
12	1931	July	8.99	1671
13	1942	July	8.23	1104
14	1942	October	8.84	1557
15	1945	June	8.61	1388
16	1946	April	9.45	2039
17	1949	June	11.15	4106
18	1950	February	9.60	2166
19	1950	April	8.84	1557
20	1950	June	10.82	2534
21	1950	June	10.36	2095
22	1950	June	8.53	1331
23	1950	July	8.69	1076
24	1950	November	9.14	1331
25	1951	January	10.97	2690
26	1952	August	11.51	3540
27	1952	August	11.20	3228
28	1953	May	9.45	1600
29	1954	February	9.60	1685
30	1955	February	12.31	10336
31	1956	March	8.46	1648
32	1956	June	8.61	1730
33	1962	May	10.59	2662
34	1964	June	10.59	2662
35	1967	August	8.89	1390

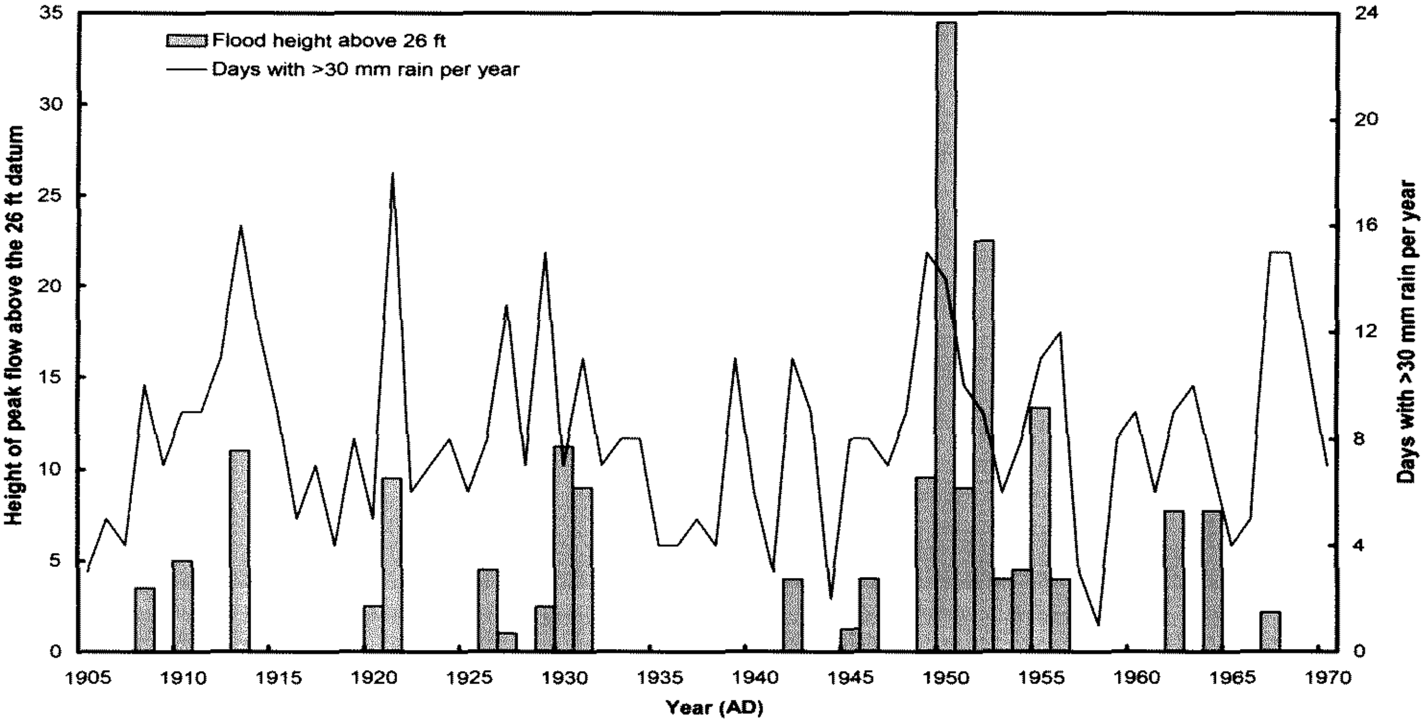


Figure 2.14 Number of days per year that the 30 mm initial storm loss was exceeded in the middle Paterson River valley, plotted against the height above 26 feet (floods are defined as occurring when the stage equals or exceeds 27 feet) that flood waters reached at Belmore Bridge, Maitland between 1908 and 1967. Flood observations taken from Anon. (1968b).

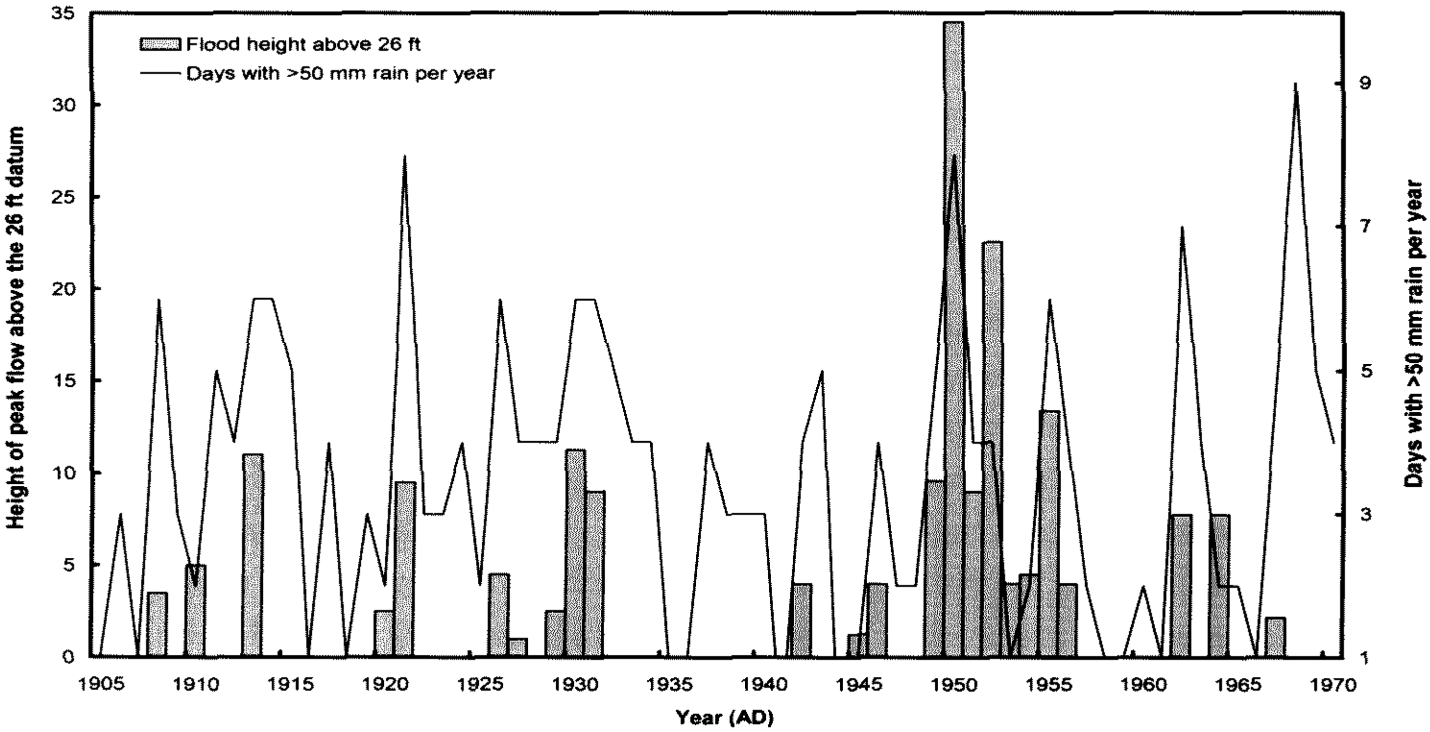


Figure 2.15 Number of days per year that the 50 mm initial storm loss was exceeded in the middle Paterson River valley, plotted against the height above 26 feet (floods are defined as occurring when the stage equals or exceeds 27 feet) that flood waters reached at Belmore Bridge, Maitland between 1908 and 1967. Flood observations taken from Anon. (1968b).

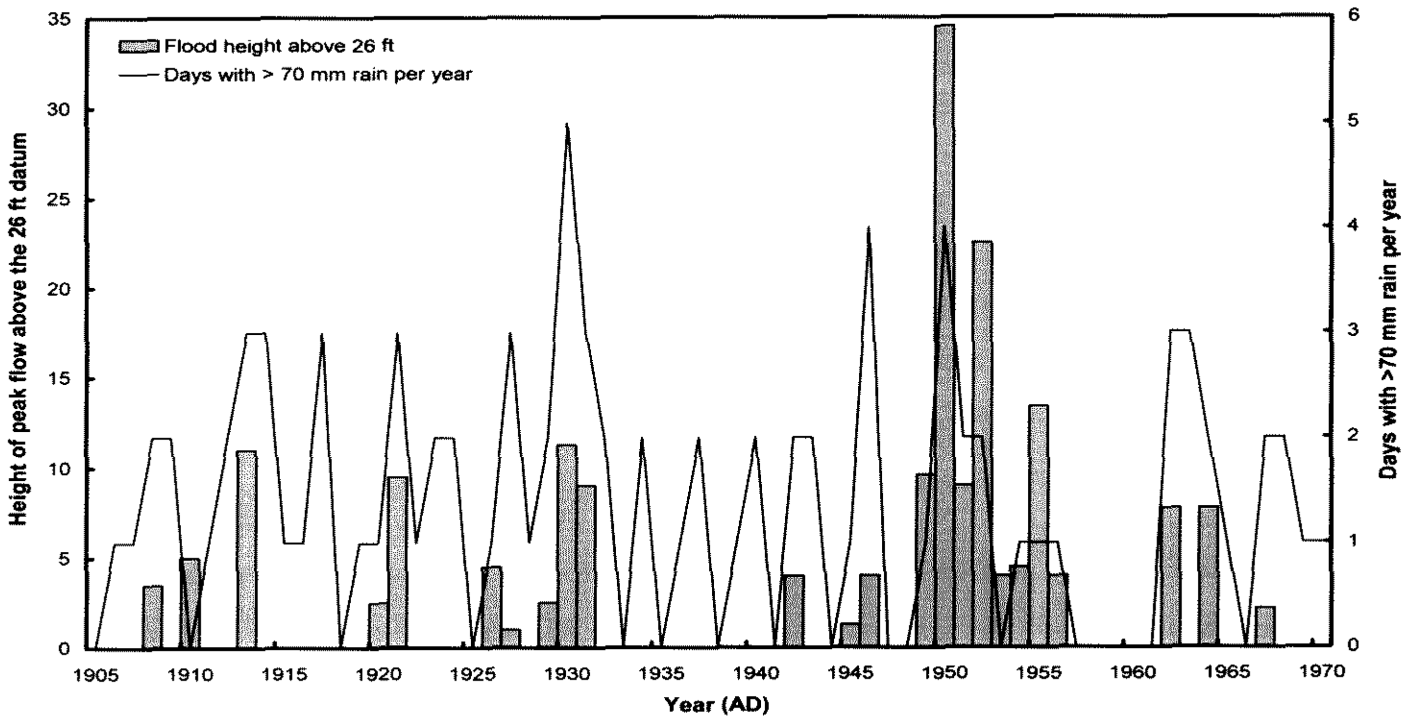


Figure 2.16 Number of days per year that the 70 mm initial storm loss was exceeded in the middle Paterson River valley, plotted against the height above 26 feet (floods are defined as occurring when the stage equals or exceeds 27 feet) that flood waters reached at Belmore Bridge, Maitland between 1908 and 1967. Flood observations taken from Anon. (1968b).

The lower limit of the initial storm loss estimate (>30 mm rain) exhibits a broad fit with the observed floods (Figure 2.14). Typically, years that experience more than five days with >30 mm rainfall recorded a flood at the Belmore Bridge gauge at Maitland. The relationship between the two is not, however, always synchronous. In the late 1910s, from 1932 to 1940, and in the early 1960s, the number of days per year that exceeded 30 mm rain in the Paterson River valley ranged between 5 and 10, with no floods recorded.

The initial loss model of >50 mm appears to be a good predictor of the larger floods at Maitland, particularly those years that had multiple floods (Figure 2.15). Importantly, there also seems to be some relationship between the magnitudes of the two records, such that the largest number of >50 mm rain days corresponds to the year that recorded the highest number of floods at Maitland (1950). Smaller floods also occurred in years with higher numbers of days per annum with >50 mm rain. However, there are several periods when increased rainfall in the Paterson valley is decoupled from floods at Maitland. These include the late 1910s, 1923–1924 and 1932–1940. The large increase in the number of >50 mm rain days in the late 1960s is beyond the period of flood observations recorded in Anon. (1968b). Overall, this model is more successful in representing flood events than that of Figure 2.14.

The initial loss model of >70 mm (Figure 2.16) appears to be in phase with every major flood recorded at the Belmore Bridge gauge. Those years when the Hunter flooded at Maitland generally experienced a minimum of three days of >70 mm rainfall in the Paterson valley. The single exception to this is the year 1910, when a moderate flood was recorded, yet no >70 mm rainstorms were recorded in the Tocal region. As with the >30 and >50 mm initial storm loss models, the late 1910s and the period from 1932 to 1940 were characterised by no flooding, and yet several years during these periods recorded more than five days of >70 mm rainfall. Unlike the other storm loss models, the >70 mm model mirrors well the flood behaviour of the Hunter River at Maitland in the early 1960s.

2.6.3 Summary

While all three models provide reasonable approximations of the flood conditions of the Hunter River at Maitland through the 20th century, the initial working storm loss model of >50 mm and the upper limit model of >70 mm were more successful. There were, however, several extended periods of heavy storm losses in the Paterson River valley when no flooding was recorded at Maitland, notably the late 1910s, and from 1932 to 1940. That some disconnection exists between these two types of hydrological record is not surprising. The Hunter River at the Belmore Bridge, Maitland is fed from the much larger upstream catchment of the Hunter valley. The catchment area upstream of Maitland is an order of magnitude greater than the Paterson River valley, and the rivers drain geomorphically and climatically contrasting regions.

As these storm models have demonstrated, heavy rainfalls in the Paterson valley will not always result in flooding downstream at Maitland, and vice versa. Nevertheless, the >50 mm and >70 mm storm loss models broadly mirror the history of flooding at Maitland, which suggests that in years when significant rainstorms flooded the Paterson River valley, similar rainfalls were recorded in the upper Hunter valley leading to flooding at Maitland. When years dominated by heavier rainfall were confined to the Tocal region alone, these data suggest that corresponding rises on the Hunter at Maitland may not have taken place. This work also suggests that the storm loss modelling probably provides a better proxy of years of wetter conditions in the Tocal region than stage data from nearby rivers. In any case, rainfall records in the Hunter valley significantly predate flood stage data, thus a longer proxy of past flood or drought conditions may be obtained.

Chapter Three

The late Holocene in eastern Australia

3.1 Introduction

This chapter aims to review our understanding of environmental change in eastern Australia during the Holocene. It will focus particularly on the temperate environments of central eastern Australia. The impact of both Aboriginal people and European colonists is considered, and the knowledge we have of the imprint of human activities on the environment is reviewed.

In the first instance, this review shall focus on those published records of palaeoenvironmental significance that have a sound chronological basis. In addition, while well-dated records have been obtained from a variety of environments, this review is biased towards lake-sediment based reconstructions, in part because this complements the work reported here from Tocal. In most instances, the original interpretation of the data has been presented, but in cases where uncertainty or disagreement exists, then this is discussed. A map of eastern Australia is shown in Figure 3.1, giving the locations of major cities and other features referred to here. Figure 3.2 shows the locations of those sites discussed in the chapter.

Much of the chronology for these studies has been based on ^{14}C methods. As the use of both uncalibrated and calibrated ^{14}C ages is common in much of this work, I have reported ages in calibrated calendar years BP (abbreviated to cal. BP) in an attempt to standardise the data sets, with uncalibrated ages converted using CalPal (the Cologne Radiocarbon Calibration and Paleoclimate Research Package). Finally, dates from the last 1000 years are expressed in calendar years (AD) (conversions made using CalPal), to allow comparison with the prevalence of non-^{14}C ages in many studies of this period.

A much more detailed review is presented of environmental change in the last 1000 years, with particular reference to those changes that have taken place in response to the arrival of Europeans in Australia. Some background to this topic is provided by examining the environmental impacts that followed European colonisation and the onset of modern agriculture in other parts of the world. This review is made all the more relevant considering the small number of studies from the Australian continent that have addressed this topic, and the lack of consensus that plagues this research (Butzer and Helgren, 2005). Lake-sediment based records of catchment erosion are discussed, reviewing the Australian literature of lake-based sediment yield research. Finally, those few records that have extended the lake record of catchment soil erosion back to before the time of European settlement are discussed. Such records have provided the only quantitative information to date on the timing and magnitude of European impacts on the Australian environment. The historical record of European settlement on Australian ecosystems is then addressed, along with its relationship to the palaeoenvironmental record.

In summary, this chapter seeks to detail the Holocene-scale environmental backdrop against which human modification of the Australian environment has taken place, while also noting where gaps in our knowledge currently exist. This is of great importance to studies such as this, where human modification of the environment may have been contemporary with climatic shifts, and therefore where the relative importance of each factor must be evaluated.

3.2 Holocene environmental change in eastern Australia

3.2.1 Introduction

Although the environmental changes that have take place during the Holocene are often considered relatively small when compared with those experienced through glacial/interglacial cycles, a solid understanding of Holocene environmental shifts and of their geographical distribution is essential for the study of human impact in Australia. Confusion regarding the timing and magnitude of climatic impacts on the environment through the Holocene (such as around the time of first European settlement in Australia [Tibby, 2003]), has the potential to distort our understanding of the impacts of anthropogenic activities on the environment.

Most Australian palaeoclimatic records and most published lake-sediment data are from the temperate regions of the continent, and mainly from the southeast coastal fringe. The focus on this relatively small part of Australia is largely a consequence of the existence of sites with the potential for the preservation of environmental records, rather than any singular importance of the region at the expense of the remainder of the continent. Some of this bias may also reflect the concentration of Australia's population and researchers in the region. It is these records that will be considered mostly here, as they are of most relevance to the pattern of Holocene climatic change thought to have been experienced at Tocal Homestead Lagoon. As demonstrated by Harrison's (1993) synthesis of Late Quaternary lake-level changes in Australia, climate in the Late Quaternary has varied both spatially and temporally, with the main differences being between the tropical and temperate regions of the continent, and between inland and coastal regions.

Past climatic conditions during the Holocene in Australia have traditionally been inferred from pollen sequences in lake sediments, though a few important (and often high-resolution) records have come from geomorphological, tree-ring, speleothem and coral studies. A consequence of the dominant use of pollen records from lakes in this region has been a limitation in the type of palaeoenvironmental information typically obtained. Some changes in vegetation types (and thus, shifts in pollen assemblages) may be driven by changes in effective precipitation (EP). These records have thus been used as indicators of wetter or drier climates. By establishing bioclimatic links (Kershaw *et al.*, 1994), other Australian pollen records have been used to infer changes in temperature, and even cloud cover. Shifts in pollen records, however, may equally be the result of soil processes, time and human activity. Changes in pollen assemblages that have been explained away by climate alone may potentially overestimate the importance of its role in the Holocene. This is a particular concern in the light of the suggested insensitivity of Australian vegetation to sub-Holocene scale environmental shifts (Dodson *et al.*, 1986).

Fossil charcoal records from lakes, often obtained in conjunction with pollen records, have provided indices of past fire frequencies and intensities (Kershaw, 1986). These palaeo-fire records have played an important role in debates about Aboriginal land use and fire practice before European settlement (Mooney *et al.*, 2001), and have provided information on the debated antiquity of Aboriginal occupation in Australia (Singh *et al.* ,1981) (despite a human origin of past fires rarely being established with any certainty).

Australian tree-ring records have been used to provide reconstructions of warm-season surface air temperatures (Cook *et al.*, 1992; 2000). Speleothems too retain information on past surface air-temperatures (Goede *et al.*, 1996) and rainfall (Fischer, 2004), while coral retains records of surface sea temperatures (Gagan *et al.* 2004) and, in some instances, records of sediment flux to the ocean (McCulloch *et al.*, 2003).

3.2.2 Holocene records from central eastern Australia

Only a small number of palaeoenvironmental records has been published to date from sites in the Hunter Valley region of eastern Australia. As a result, this review will also focus on sites from throughout central eastern Australia. Most Hunter valley records are from Barrington Tops, a high-altitude (1000 to 1500 m ASL) plateau located approximately 60 km northeast of Tocal Homestead Lagoon (Figure 3.2). From the plateau, fossil pollen and charcoal records (and one sediment magnetism record) have been obtained from a series of swamps. These are mostly of Holocene age, although one record extends back to 40 000 cal. BP (Sweller and Martin, 2001). Although these sites are close to Tocal

Homestead Lagoon, they experience a climate very different to that of Tocal, and the environmental record from Barrington cannot be simply extrapolated to lowland Hunter Valley sites. At the southeast edge of the Hunter Valley, Redhead Lagoon possesses a palaeoenvironmental record that spans the last 70 000 years (Williams, 2005). The environment of this site is more closely comparable with that of Tocal.

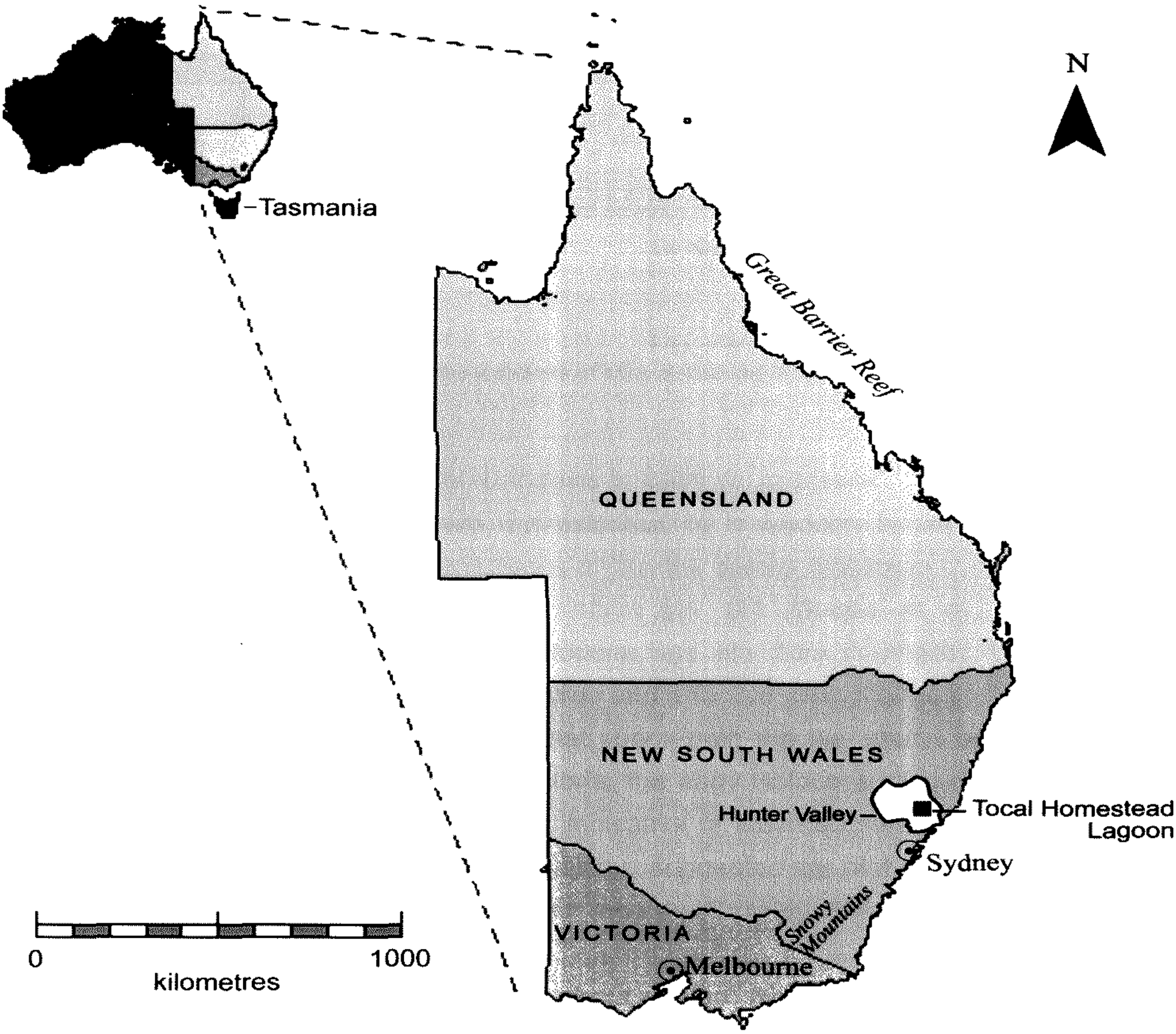


Figure 3.1 The eastern states of Australia, showing the location of capital cities and major features referred to in this chapter.

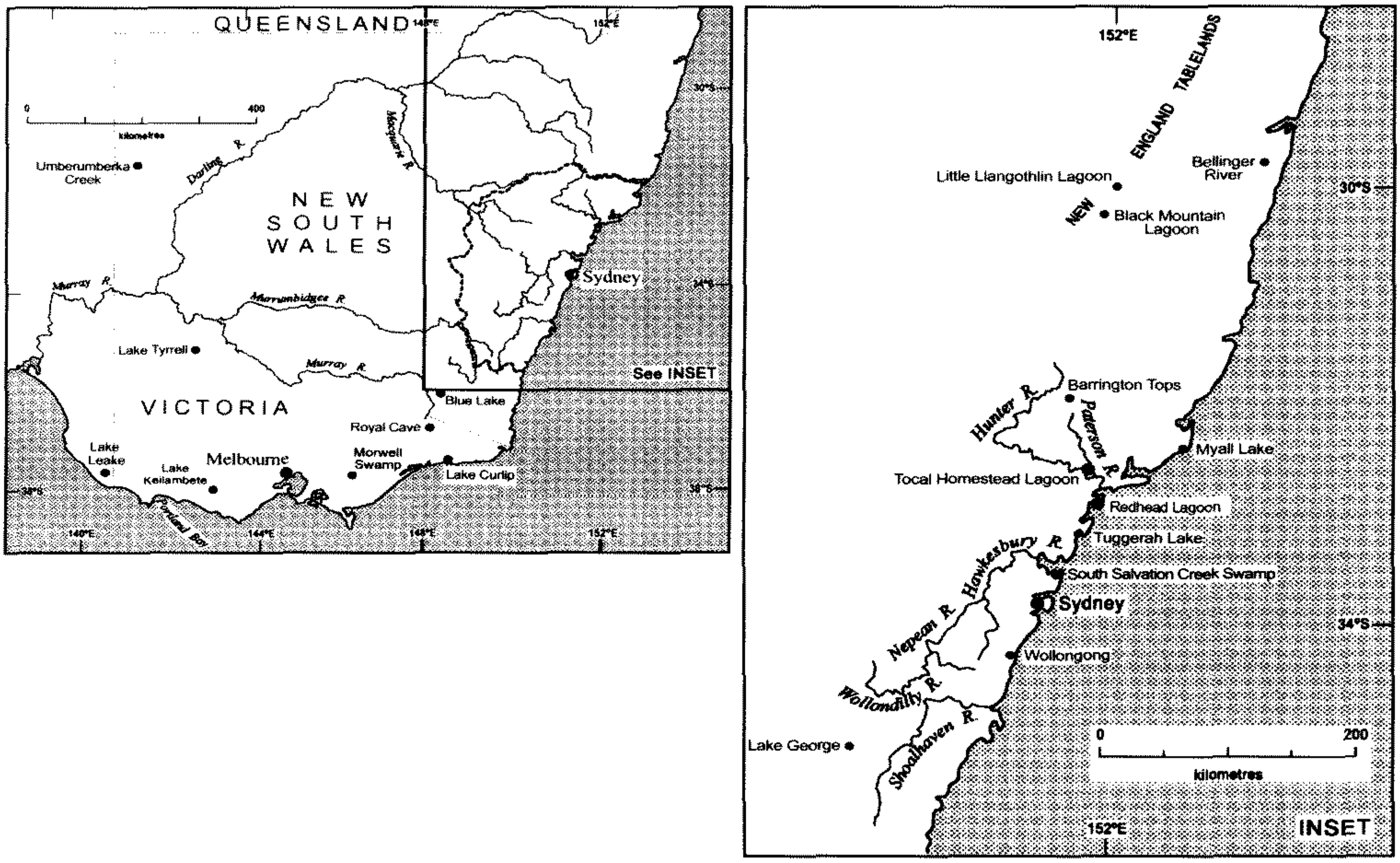


Figure 3.2 Southeast Australia, showing the locations of sites discussed in this chapter.

3.2.3 Climate of the early Holocene in eastern Australia

The early Holocene in Australia was characterised by a recovery to warmer and wetter conditions from the cool and arid conditions of the last glacial maximum *c.* 17 000 cal. BP and the Younger Dryas of 13 000 to 11 500 cal. BP (Goede *et al.*, 1996). Most palaeoenvironmental records of early Holocene age are from southeast Australia, and generally suggest an increase in temperatures and effective precipitation through this time (Lloyd and Kershaw, 1997). In a record that spans both the last glacial and post-glacial periods from Lake Tyrrell in northwest Victoria, the early Holocene recorded increases in grass pollen and lake levels (Luly, 1993), indicative of elevated summer rainfall. Wetter conditions are paralleled by increasing salinity, suggested by higher concentrations of salt-tolerant species in early Holocene pollen records (Crowley, 1994).

Analysis of lake-level data sets from throughout eastern Australia shows both a tropical-temperate variation and a coastal-interior variation in this pattern of climatic amelioration. Some Tasmanian lakes recording elevated lake levels as early as 11 000 cal. BP (Harrison, 1993), while Goede *et al.*'s (1990) Tasmanian speleothem records suggest temperatures peaked in the early Holocene and exceeded those of today. One site from

Harrison's (1993) compilation, Lake Keilambete, southern Victoria (Bowler, 1981), provides evidence suggesting that lake levels were at their lowest point during the Holocene around 11 000 cal. BP, but began to increase sharply after this time, particularly after 9600 cal. BP (Figure 3.3). This site also records a short-lived maximum in lake levels around 9200 cal. BP, an event not identified anywhere else to date. Pollen records from southeast Australian lakes suggest that rainforest communities were probably more extensive in the early Holocene than at present, indicating that wetter conditions than present may have prevailed (Harrison, 1993). While most sites suggest increasingly wetter conditions in the early Holocene, temperatures during this period appear to show some localised variations. At Morwell Swamp in southeast Victoria, a quantitative estimate of early Holocene conditions has suggested that summer temperatures were at least 1.3°C warmer than today around 9000 cal. BP (Lloyd and Kershaw, 1997). The stalagmite palaeoclimate history from Royal Cave, Buchan, eastern Victoria (Goede *et al.*, 1996), however, suggests that from 9000 to 8000 years ago temperatures were approximately 2°C cooler than present.

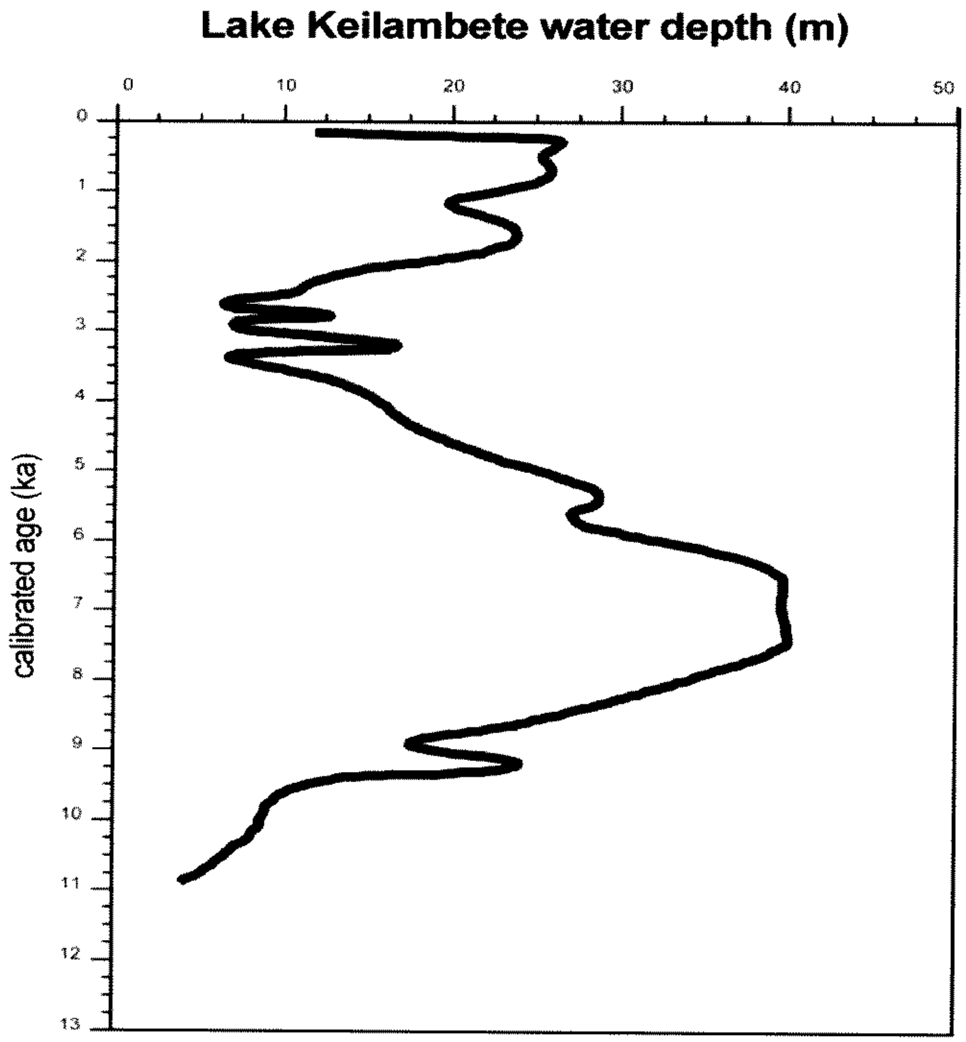


Figure 3.3 The Lake Keilambete water level curve, adapted from Bowler (1981). The vertical (age) axis is in calibrated years BP.

3.2.4 The early Holocene in central eastern Australia

Very little direct information is available on environmental conditions in the early Holocene either in the Hunter valley or throughout central eastern Australia. Those few swamps from Barrington Tops whose records pre-date the mid-Holocene were accumulating mineral sediment before *c.* 9000 cal. BP (Dodson *et al.*, 1986), probably reflecting the end of glacial period aridity. All the main vegetation types found today had been established on Barrington Tops by *c.* 9000 cal. BP (Sweller and Martin, 2001). Forest vegetation increased through the early Holocene, with *Nothofagus* pollen increasing from *c.* 9000 cal. BP (Dodson *et al.*, 1986, Sweller and Martin, 2001), probably reflecting climatic amelioration during this period.

At sites on the Colo River, in the Hawkesbury River valley of eastern New South Wales (Figure 3.2) mangrove and salt marsh vegetation grew from *c.* 8000 cal. BP to *c.* 6000 cal. BP (Jones and Dodson, 1997). As sea-level on the east coast of Australia did not reach modern levels until *c.* 6500 cal. BP (Thom and Roy, 1983), marine water incursions up into the valley cannot explain early Holocene saline conditions. That salt-tolerant vegetation grew *c.* 2000 years before maximum sea-levels were attained may reflect increased salinity in the region in the early Holocene, related to higher ground water tables and wetter conditions. Similar early Holocene increases in salinity throughout temperate southeast Australia have been related to wetter conditions (Crowley, 1994).

3.2.5 Climate of the mid-Holocene in eastern Australia

Peak temperatures and maximum lake levels (and by extension, maximum effective precipitation), were recorded between 6000 and 4000 cal. BP at many sites in southeast Australia (Head, 1988, Harrison and Dodson, 1993, Kershaw, 1995, Gell, 1997). These inferences have generally been drawn from pollen changes in lake sediment records, particularly from increases in wet sclerophyll and rainforest taxa to a maximum in the mid-Holocene (e.g., De Deckker *et al.*, 1988). Pollen, diatom and lake level studies suggest that temperatures were as much as 1°C higher than present, and that effective precipitation was possibly 30% greater (Dodson and Mooney, 2002).

Based on the proxy aeolian activity record from Blue Lake in the Snowy Mountains of southeast Australia, wind strengths or storm frequencies from 7000 to 5000 cal. BP were the lowest recorded for the entire Holocene (Stanley and De Deckker, 2002). These results are reflected in lake levels from southeast Australia, which reached a maximum by *c.* 6000 cal. BP (Harrison and Dodson, 1993). A significant increase in *Callitris* pollen in the sediments of Lake Tyrell, western Victoria after 6600 cal. BP led Luly (1993) to suggest that rainfall at the time was as much as 2.5 times higher than present.

In the Australasian region, it has been proposed that the onset of modern El Niño–Southern Oscillation (ENSO) periodicity 5000–4000 cal. BP (Shulmeister and Lees, 1995; Gagan *et al.*, 2004) initiated present-day oceanic circulation and climatic patterns in the Australian region (Allan and Lindesay, 1998). Dodson *et al.* (1986) have suggested that temperature peaked with precipitation during this period, while other pollen records (McKenzie and Busby, 1992) indicate that summer temperatures were lower than those of the present and winter temperatures slightly higher at this time. Both interpretations, however, may simply be artefacts of using vegetation alone as a climatic surrogate. Arguably the highest resolution record from this period is from Lynd's Cave in northern Tasmania (Xia *et al.*, 2001). This speleothem record suggests that maximum temperatures in Tasmania date from 8000 to 7400 cal. BP, with peak wet conditions not occurring until afterwards, from 7400 to 7000 cal. BP. In the light of these findings, the warmer and wetter conditions suggested by shifts in fossil pollen from the mid-Holocene in Australia are likely to represent separate episodes, which the vegetation record has blurred together.

In summary, the mid-Holocene in temperate eastern Australia was characterised by a peak in temperature and effective precipitation and depressed aeolian activity. The wettest conditions occurred in Tasmania and southern mainland Australia between about 7000 and 4000 cal. BP, and between 5000 and 3700 cal. BP in northeastern Australia (Shulmeister *et al.*, 2004). Several authors have argued that this temporal and spatial gradient suggests a distinct regionality accompanied environmental change in the mid-Holocene, while Shulmeister (1999) has proposed that a southward extension of the northern Australian monsoon may have led to increased summer rainfall in the mid-Holocene. This interpretation may be in part influenced by a reliance on pollen data. By contrast, corals from tropical north Queensland record mid-Holocene maximum sea-surface temperatures 1.2°C higher than those of the 1990s AD in *c.* 5400 cal. BP (Gagan *et al.*, 1998), coincident with maximum temperatures obtained at some sites at higher latitudes.

3.2.6 The mid-Holocene in central eastern Australia

Well-dated evidence from the coast of eastern Australia points to warmer conditions than present in the mid-Holocene (Baker *et al.*, 2001; Haworth *et al.*, 2004). The spatial and temporal distribution of the Dugong (*Dungong dugon*) fossil record has provided solid evidence of warmer climates and higher sea levels than those of the present in the Sydney region during the mid-Holocene (6320 to 5920 cal. BP). Relict subtropical tubeworms in Port Hacking, southern Sydney (Haworth *et al.*, 2004) dated to the mid-

Holocene suggest that as late as *c.* 4000 cal. BP ocean waters off Sydney were up to 2°C warmer than at present. Interestingly, Baker *et al.* (2001) suggested that a brief period of colder ocean temperatures occurred around *c.* 5100 cal. BP, bisecting the 'warm period' identified for the mid-Holocene in other studies. However, such variability might be related to the complexity of the mathematical curves fitted to their biological indicator data.

Fossil subtropical corals from the Wollongong region of central eastern Australia have been dated to the mid-Holocene using U/Th methods (Young *et al.*, 1993). These corals are today found at least 500 km north of Wollongong. Their presence implies sea temperatures in the region were of the order of 2.5°C warmer than present. However, two other U/Th dates on these corals are significantly younger (2800 and 2000 cal. BP), suggesting that comparable conditions existed at some stage in the late Holocene.

Evidence of wetter conditions in the mid-Holocene comes from sediment cores from the Hawkesbury River valley, immediately south of the Hunter valley (Devoy *et al.*, 1994). These suggest that despite rising sea-levels, and thus inundation of the valley after *c.* 8000 cal. BP, there was an increase in fluvial infill and freshening of the water is evident around the mid-Holocene. Sedimentological evidence from the Sydney region suggests that wetter conditions in the mid-Holocene may have ceased by *c.* 4000 cal. BP, with increased coastal dune instability in Sydney dating from this time (Martin, 1994). The destruction of dune forest in Fingal Bay, Port Stephens, approximately 70 km east of Tocal, occurred between *c.* 6000 and *c.* 3000 cal. BP (Macphail, 1974), and is thought to be related to rising water-tables in response to warmer mid-Holocene conditions.

Pollen records from Barrington Tops suggest that the mid-Holocene was perhaps 1-2°C warmer than at present (Dodson *et al.*, 1994). Peat formation began at Burraga Swamp about 6500 cal. BP, after almost 35 000 years of predominantly minerogenic deposition (Sweller and Martin, 2001). Curiously, while other sites on the Barrington Tops plateau had entered a phase of peat formation by 3000 cal. BP (Dodson *et al.*, 1986), others did not complete the shift until much more recently (740 cal. BP) (Dodson, 1987). These differences have been related to elevation and to the effects of aspect on rainfall. However, the lack of synchronicity in peat formation on Barrington Tops may also be indicative of the minor role played by climate.

In the mid-Holocene, *Nothofagus* forests reached full development around Burraga Swamp on Barrington Tops, (Sweller and Martin, 2001), perhaps in response to wetter conditions. Maxima in mineral sediment flux (and, by extension, catchment erosion) were recorded in the mid-Holocene in Black Mountain Lagoon and Redhead Lagoon, eastern New South Wales. These sedimentation regimes have been interpreted as climatically driven (Gale *et al.*, 2005), with the Redhead record providing some pollen-based evidence of a higher water balance during this period (Williams, 2005).

3.2.7 The late Holocene in eastern Australia

Many authors (e.g., McGlone *et al.*, 1992) have suggested that the late Holocene in Australia was characterised by frequent climatic fluctuations. This proposed environmental variability may be due in part to the increased resolution of many records in eastern Australia from this period. Tree-ring records of annual resolution have recorded frequent, albeit often minor, temperature fluctuations in Tasmania (Cook *et al.*, 1991; 1992), while coral-based records from northeast Queensland have provided reconstructions of mean annual sea-surface temperatures, and salinities, rainfall and terrestrial runoff at an annual and sub-annual resolution (Hendy *et al.*, 2002; Gagan *et al.*, 2004; Lough, 2004; Gagan, 2005). Unfortunately, no published records of this quality exist for the remainder of the Holocene in eastern Australia. It is also possible that the increase in environmental variability is related to the intensification of El Niño–Southern Oscillation climatic processes after 5000 cal. BP (Hayne and Chappell, 2001), leading to enhanced decadal and centennial scale fluctuations between extreme wet and dry periods (Stanley and De Deckker, 2002).

Pollen records from southeast Australia generally record a return from wetter conditions sometime after 5000 cal. BP. McKenzie and Busby's (1992) data for the late Holocene suggest a temperature increase at about this time. The lake level curve from Lake Keilambete in western Victoria (Bowler, 1981) records a water-balance minimum around 3500 cal. BP, with slightly higher lake levels recorded later in the Holocene. A diatom-based record from Lake Curlip, eastern Victoria shows an increase in salinity from *c.* 3500 to *c.* 2200 cal. BP, which may be in response to increased aridity in the region from this time (MacGregor *et al.*, 2005). Evidence of increased salinity in the Lake Leake record is also thought to be related to aridity in southeast Australia from *c.* 3000 to *c.* 2000 cal. BP (Crowley, 1994). Nearby, the speleothem-based palaeotemperature record from Royal Cave, eastern Victoria (Goede *et al.*, 1996) shows a marked cool event dating from about 3000 cal. BP, when temperatures were estimated to be as much as 2 to 3°C lower than at present. This supports earlier work from montane areas in southeast Australia that proposed a decrease in temperature in the late Holocene (Costin, 1972; Macphail and Hope, 1985). Not all records are in agreement for this period, however, with evidence of wetter conditions at Lake George, southeast New South Wales from *c.* 3200 cal. BP

(Singh and Geissler, 1985). Overall, sites in southeast Australia suggest that drier and cooler conditions prevailed from some time between 3500 and 3000 cal. BP until *c.* 2000 cal. BP.

Mooney's (1997) high-resolution record from Lake Keilambete provides evidence of increased effective precipitation from 2000 to 1800 cal. BP, with Dodson's (1974) pollen record from Lake Leake displaying similar changes for this period. Decreased moisture and cooler temperatures from 1750 to 1430 cal. BP have been inferred from the Keilambete record, while from 1420 to 1320 cal. BP the region's climate may have been characterised by either warmer temperatures or increased effective precipitation (Mooney, 1997). Most pollen and charcoal records from southeast Australia suggest that fires became more common in the late Holocene (Kershaw *et al.*, 2002), which may relate either to hypothesised increases in Aboriginal population numbers during this time (e.g., Lourandos, 1997), to drier conditions, or perhaps both. Moving further inland, lake records show greater aridity after *c.* 4000 cal. BP (Cupper, 2005), with increased dune instability also characteristic of this period (Wasson, 1982).

Not all late Holocene environmental change has been attributed to climatic shifts. Valley fills from eastern Australia that date from the last 3000 to 4000 years have been interpreted as the result of increased Aboriginal burning of the environment (Hickin and Page, 1971; Hughes and Sullivan, 1981), with Kohen (1995) concluding '... that it is difficult to find any other satisfactory explanation ...'. However, with little data and no consensus on the climatic history of the mid- to late Holocene in eastern Australia, a climatic origin of these features is equally, if not more likely than a sudden rise in human activity capable of major geomorphological change. Williams (1978) dated hillslope instability and stream aggradation on the Southern Tablelands of inland southern New South Wales to between 4500 and 1400 cal. BP. He suggested that this could be linked to lower temperatures and to drier and windier conditions in the late Holocene rather than to human impact. These changes in stream aggradation and gullying have also been explained by Prosser (1990) in the context of the internal dynamics of geomorphic system behaviour. This interpretation was based predominantly on a lack of synchronicity in gullying events in catchments on the Southern Tablelands (Prosser, 1990; Prosser et al., 1994; Gillespie *et al.*, 1992).

Climatic instability in the Lake Keilambete region from around 1750 to 1400 cal. BP was identified from increases in minerogenic matter and parallel decreases in organic material in the sediment sequences (Mooney, 1997). These may have been the result of decreased moisture content and/or temperatures during this period. While other climatic changes, or even human impact, could explain these changes in the Keilambete record, some support for Mooney's (1997) interpretation of drier conditions during this time comes from lowered water levels in salt lakes in the region (Bowler, 1981). However, the Lake Keilambete lake level curve displays some evidence of higher lake levels from around 1800 to 1300 cal. BP years (Figure 3.3). Dodson *et al.* (1986) suggested that no significant climatic changes have occurred within the last 1500 years. It is far more probable, however, that any such changes are simply of lower magnitude than those of the rest of the Holocene, and that the strength of their signal is below the sensitivity, or chronological resolution, of the predominantly pollen-based records published to date.

3.2.8 The late Holocene in central eastern Australia

Changes in swamp and terrestrial pollen in South Salvation Creek swamp, northern Sydney, have been interpreted as indicating drier climates from *c.* 2000 cal. BP to the present. Fire frequencies at this site appear to increase after the mid-Holocene, reaching a peak at *c.* 2000 cal. BP (Kodela and Dodson, 1988). This may reflect an intensification Aboriginal land use during the late Holocene (Kodela and Dodson, 1988).

The pollen record from Burraga Swamp, Barrington Tops, shows only minor changes during the last 2000 years, except for some evidence of cooler conditions from around 1500 cal. BP (Dodson *et al.*, 1986; Dodson, 1987). A range of well-dated sediment cores from other swamps on the Barrington Tops shows a shift from minerogenic to peat accumulation mainly around 3000 cal. BP. At lower altitude sites, however, this change began in the mid-Holocene (section 3.2.6). These changes were accompanied by declines in rainforest pollen, which were accounted for by a suggested lowering of temperatures and precipitation from this time (Dodson, 1987). As no corresponding increase in peatland has been recorded in the similar region of Tasmania, Dodson (1987) proposed that the late Holocene expansion of peat may be limited to lower latitude subalpine sites only, possibly resulting from greater penetration of summer rain-bearing systems during this period.

Late Holocene increases in herbaceous taxa in the pollen record from Redhead Lagoon have been interpreted by Williams (2005) as possibly recording climatic decline. A rapid decrease in minerogenic input to Terragong Swamp is very clearly marked, and dates from *c.* 2500 cal. BP (Jones, 1990). The shift in sediment type from terrestrial peat to marine mud deposition was interpreted by Jones (1990) as probably recording the stabilisation of sea-level near the present-day level.

3.3 Environmental change in the last 1000 years in eastern Australia

3.3.1 Introduction

Overwhelmingly, the greatest cause of environmental modification across the Earth during the last 1000 years has been human activity (Roberts, 1998). It is pertinent to note, however, that significant global-scale climatic variability has also been recorded during this same period. Without some understanding of the role of sub-millennial scale climatic fluctuations in forcing environmental change, the impact of anthropogenic activities may be obscured or misunderstood. Disentangling the role of these two processes in forcing environmental change remains difficult (Brauer and Guilizzoni, 2004), while many studies of recent environmental change do not consider the role of climate at all. As Wilby *et al.* (1997) have noted, when sedimentary records of recent environmental change are examined, anthropogenic processes are often invoked to explain the entire record, with little or no consideration given to the role of climatic variability. The task of explanation is made even more difficult by the complexity of the interaction of human and climatic processes. While at earlier times, environmental change influenced human activity, during the last 2000 years, and particularly since the Industrial Revolution, the interrelationship between the two has grown in complexity. Theophrastus of Eresus in Lesbos, a student of Aristotle, suggested as early as the 4th century BC that deforestation and landscape modification may have been responsible for climatic change (Glacken, 1967). In recent decades, the keenly debated connection between the anthropogenically increased CO_2 levels in the atmosphere and climatic change demonstrates the geopolitical importance of the interaction between climatic and human processes.

Research from the northern hemisphere has produced a large database of palaeoclimatic information for the last 1000 years. Much of this work has distinguished two main periods of contrasting climatic conditions; the Medieval Warm Period and the Little Ice Age. The Medieval Warm Period is thought to have begun around the 9th century and continued into the 14th century (Hughes and Diaz, 1994), while the Little Ice Age is considered to have begun in the 13th or 14th centuries and culminated between the 16th and 19th centuries (Jones and Bradley, 1992; Grove, 2004). These episodes may have occurred asynchronously in different parts of the world (Goosse *et al.*, 2003). Whether there is any evidence for the occurrence of these events in Australian is yet to be established.

3.3.2 Climate of the last 1000 years in eastern Australia

The paucity of sub-millennial scale palaeoenvironmental records from Australia has meant that only a few sites have provided information on this period. The annual resolution of these few records, mainly from tree-rings or corals (Cook *et al.*, 1991; 1992; 2002; Gagan *et al.*, 2004) has ensured that they are, in theory, superb sources of information on environmental change in Australia for the last 1000 years. Their only real problem is the very narrow range of Australian geographical regions that they represent; usually either mid-latitude subalpine Tasmania, or low-latitude tropical northern Australia. To improve the detail of the picture of environmental change during the last 1000 years, important records from the greater Australasian region (mainly New Zealand and Antarctica) will be considered here as well. As much of the sub-millennial palaeoenvironmental record from Australia is not based on ^{14}C ages, dates discussed in this section will be in calendar years AD.

3.3.3 The Medieval Warm Period

The only published terrestrial record from mainland Australia that purports to have recorded climatic fluctuations during the Medieval Warm Period comes from Lake Keilambete in southeast Australia (Mooney, 1997). Several fluctuations in the organic content of the lake sediments from 800 to 1070 AD were interpreted by Mooney (1997) as a record of the Medieval Warm Period, paralleling fluctuations during this period recorded in the Northern Hemisphere by Stine (1994). Periods of warmer summer temperatures from 940 to 1000 AD, 1100 to 1190 AD and 1475 to 1495 AD have been identified in the Tasmanian tree-ring record (Cook *et al.*, 1991), with the period from 1100 to 1190 AD thought to correspond to the Medieval Warm Period in the Northern Hemisphere. The climatic signal is weak, however, and significantly less marked than the period of warming recorded after 1965 AD (Cook *et al.*, 1991). Improvements in the Huon Pine tree-ring palaeotemperature record (Cook *et al.*, 2000) have provided stronger evidence of a prolonged warm period between 900 AD and 1500 AD. Recent climatic modelling undertaken by Goosse *et al.* (2003) have also provided evidence of a 'medieval' climatic optimum in the Southern Hemisphere, which lagged behind that recorded in the Northern Hemisphere. These data, incorporating the ice core record from Laws Dome, Antarctica (Morgan, 1985), reveal the existence of warmer conditions from the 13th until the 15th century in the Southern Hemisphere, driven by heating of the Southern Ocean.

Convincing evidence for a Medieval Warm Period in temperate and alpine regions of New Zealand has been obtained from tree-ring records (Grinsted and Wilson, 1979; Cook *et al.*, 2002), speleothems (Williams *et al.*, 2004), glacial behaviour (Grove, 2004) and lake sediment records (Eden and Page, 1998). Of these, the lake and speleothem records from the temperate North Island of New Zealand may be of some relevance to eastern temperate Australia. Previous comparisons have shown that palaeoclimatic records from the northern region (30°–40°S) of New Zealand have some affinity with southeast Australian data sets (Newnham, 1999), while the region north of Auckland has been mapped as part of the same biome as temperate eastern Australia by the PEPII research program (Hope *et al.*, 2004). Finally, probably the most convincing evidence for a Medieval Warm Period in the region comes from Antarctic ice-cores. The pattern of magnetic susceptibility in an ice-core from the eastern Bransfield Basin, Antarctic Peninsula, shows distinct shifts from clastic- to organic-rich deposition, which Khim *et al.* (2002) have argued record climatic shifts synchronous with both the Medieval Warm Period and the Little Ice Age of the northern hemisphere (Figure 3.4). Their study also shows a distinct colder period from 1000 to 1100 AD indicating that brief interludes of colder conditions may have prevailed during the 'warm period'.

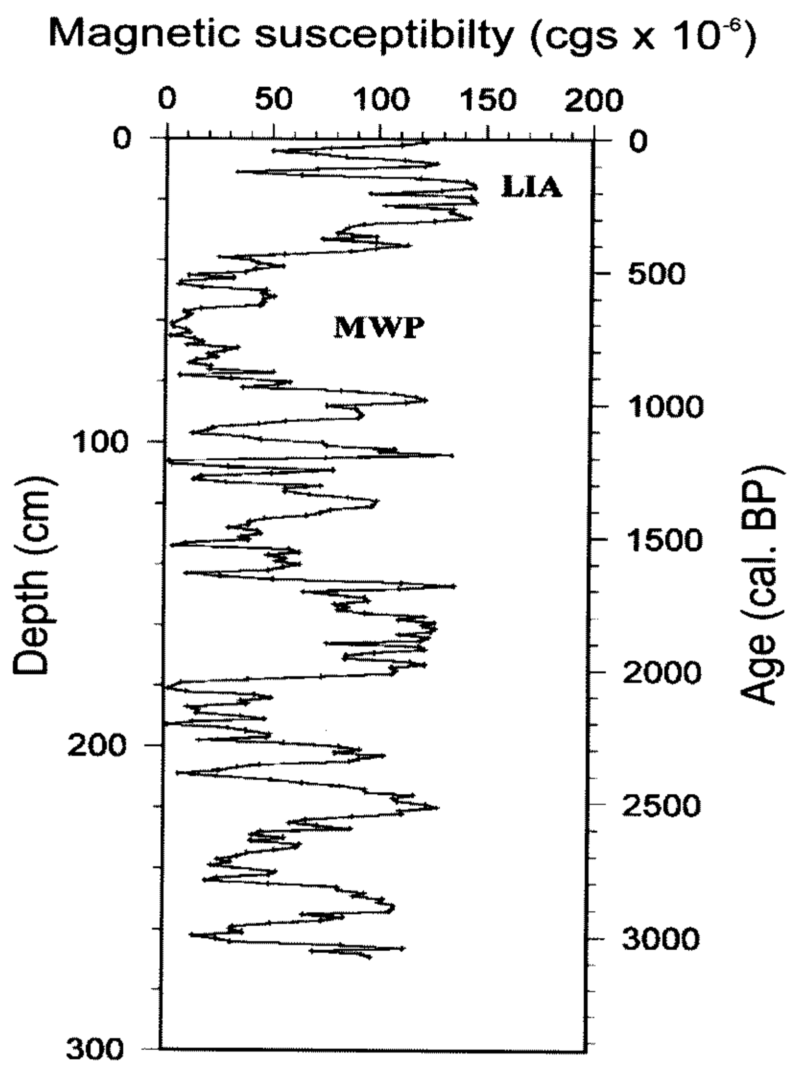


Figure 3.4 Downcore variations in magnetic susceptibility in core A9-EB2 from the eastern Bransfield Basin, Antarctic Peninsula. 'MWP' and 'LIA' are abbreviations for the Medieval Warm Period and the Little Ice Age respectively. Modified from Khim *et al.* (2002).

3.3.4 The Little Ice Age

There is little evidence in Australia for a downturn in climatic conditions in the last 500 years comparable to that recorded in the northern hemisphere and known informally as the Little Ice Age (Mann and Jones, 2003). Coral records from northern Australia, for example, suggest that the later part of this period (the 17th to 19th centuries) experienced summer temperatures comparable to those of the 1980s (Gagan *et al.*, 2004). Very few changes were detected in the Lake Keilambete sediment record for the period 400 cal. BP to *c.* 1800 AD, though the interpretation of this period is confounded by low sedimentation rates (Mooney, 1997). Jones *et al.* (2001) recorded a dry period immediately before European settlement at Lake Keilambete that may represent the end of a cooler, drier event first recorded as a dramatic drop in lake levels from the *c.* 16th century in Bowler's (1981) curve. Tree-ring records from Tasmania (Cook *et al.* 1992; 2000) show minor evidence of cooling during the Little Ice Age period, although Cook *et al.* (2000) have pointed out that their record provides little indication of a cooler period post-1500 AD (Figure 3.5).

The New Zealand palaeoclimatic database for the last 1000 years may be of some value in investigating sub-millennial climatic change in eastern Australia. This documents a period of climatic cooling corresponding to the Little Ice Age of the Northern Hemisphere (Schulmeister *et al.*, 2004; Williams *et al.*, 2004; Winkler, 2004). However, much of this evidence is related to glacial advances in the South Island of New Zealand, and may not be relevant to the changes experienced in temperate eastern Australia.

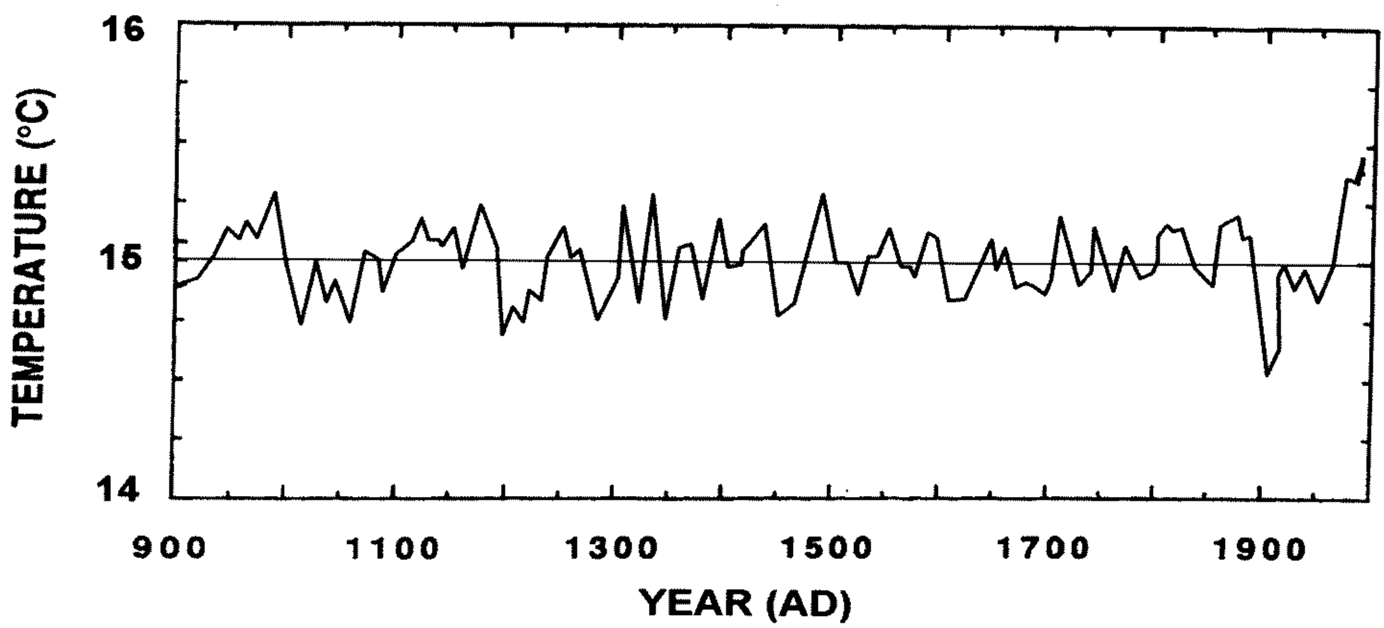


Figure 3.5 The 25-year lowpass filtered 1090-year temperature reconstruction from Huon pine for Tasmania, redrawn from Cook *et al.* (1992).

Perhaps the best evidence in the region for the occurrence of a Little Ice Age comparable to that experienced in the Northern Hemisphere comes from the ice-core records of the eastern Bransfield Basin (Khim *et al.*, 2002) and Law Dome, Antarctica (Morgan, 1985). Both records show clear periods of neoglacial conditions that equate with the Little Ice Age event. The pollen-dominated palaeoenvironmental record from Australia's temperate and semi-arid environments may not have been sensitive to these fluctuations, which may have resulted in variations in mean temperature of less than 1°C (Roberts, 1998). However, with so few data to interpret, no concrete statements can be made on climatic trends in Australia during the last 500 years before the start of the instrumental record.

3.3.5 The last 1000 years in central eastern Australia

Lake sediment records from sandstone environments to the south of the Hunter Valley appear to have recorded little environmental change during the past 1000 years (Kodela and Dodson, 1988; Dodson *et al.*, 1995), at least from a palynological perspective. The Trenerry Reserve record from Sydney (Dodson *et al.*, 1995), however, shows increasing concentrations of cation input into the wetland beginning sometime after 900 AD and before European settlement, suggestive of increased catchment erosion. However, with little chronological control, no more may be inferred from this record. In any case, a peak in Pb levels in the record long before European settlement in the late 17th century AD suggests that the geochemical stratigraphy may be compromised. By contrast, a recent study of Myall and Tuggerah lake estuaries, central eastern Australia, shows clear evidence of sub-millennial scale environmental shifts in the region at estimated 340–360 year length cycles (Skilbeck *et al.*, 2005). Unfortunately, with no dates on any of the magnetic or sedimentological features in this record younger than *c.* 1000 years, little additional information can be gained from this record.

Values of SIRM, χ_{lf} and caratenoid content decline considerably from about the 15th century AD in a sediment record from Mill Creek, in the Hawkesbury valley of central eastern Australia (Johnson, 2000). These changes were considered not to be related to human activity (such as Aboriginal burning practices), and may be the result of a climatic shift at this time, possibly to more stable (less catchment erosion) conditions. A smaller drop in cation input during this period may also be indicative of decreased erosion.

3.3.6 The period since European settlement

The end of the 18th century saw the beginning of European settlement in eastern Australia, and the environmental change associated with this is reviewed in detail in section 3.6. Beyond these changes, climatic shifts in the last 200 years have also been significant in Australia. Allan and Lindesay (1998) have summarised much of the research that has used the historical instrumental record in Australia to document periods of fluctuating rainfall, temperature and atmospheric pressure. As the earliest continuous instrumental records in Australia date from the 1830s, there remains a period of roughly half a century following first European settlement for which proxy climatic records are required. The Tasmanian tree ring record suggests that relatively little climatic change took place from 1750 to 1850 AD (Cook *et al.*, 1991). However, temperature stability cannot be automatically assumed for the rest of the Australian continent. Coral records from the Great Barrier Reef show that the period from 1700 to 1850 AD was as warm as the late 20th century, suggesting that the Northern Hemisphere 'hockey stick' temperature curve may not apply to the southwest Pacific region (Gagan *et al.*, 2005).

McCulloch *et al.* (2003) have employed a novel approach in developing a record of recent environmental change from the Burdekin River catchment in Queensland. By developing a proxy record of suspended sediment concentration in the estuarine environment (using Ba/Ca ratios of long-lived corals), They showed that periods of enhanced erosion, in 1761, 1765, 1787 and 1801, occurred prior to the arrival of Europeans in the area. Though the increases in erosion following European settlement in *c.* 1870 were in the range of five to ten times, the pre-European pulses of erosion are of significance as they are thought to have followed periods of drought, demonstrating a possible climatic control prior to European arrival.

Nicholls (1988) established a long history of connection between Australian droughts and ENSO events in South America. Zhang and Casey (1992) and Allan (1993) have correlated patterns of Australian rainfall with ENSO signatures (such as mean sea-level pressure) for much of the continent. The instrumental record shows that wetter conditions prevailed in eastern Australia from the 1850s until the early 20th century. Drier conditions dominated until the end of the 1940s, followed by a return to a wetter climate for the remainder of the century (Allan and Lindsey, 1998). These patterns are consistent with low-frequency changes in ENSO signatures, such as the *c.* 100 year record of mean sea-level pressure anomalies from Darwin (Zhang and Casey, 1992).

3.4 The impact of European settlement on the environment

3.4.1 Introduction

For many locations and societies across the world, the period of European colonial expansion resulted in the most substantial and long-lasting environmental changes experienced in the last millennium (Arnold, 1996). Beginning with the voyages of Portugal, The Netherlands, Spain, England and France in the 16th and 17th centuries, ecosystems that had experienced comparatively little human modification were suddenly accessible to Europeans. The European colonial expansion often resulted in the large-scale immigration of European people (like that experienced in North America), who brought with them Old World agricultural practices. The adoption and intensification of European agricultural technology required the clearing of native grasslands and forest, resulting in increased soil erosion and deterioration of soil properties. Histories of soil-erosion, extracted from well-dated sediment studies have contributed greatly to our understanding of the impact that European settlement had on these environments (e.g., Davis [1976]), though other lines of scientific evidence have provided much-needed data on European impacts on the environment (e.g., fire histories [Ward *et al.*, 2001]).

To date, only a few studies have successfully presented information on the environmental response to the arrival of European settlers in Australia in the late 18th century (see below). Though numerous lake-sediment based records of environmental change have now been published for continental Australia, only a handful has the chronological resolution to provide records of human activities that have taken place on decadal or centennial timescales. Considerably more attention has been paid to these questions in other parts of the world, where settler societies arrived from Europe in the last several centuries. The circumstances that prevailed following their arrival are often analogous to those experienced on the Australian mainland, and may provide insightful comparisons into how the Australian environment was modified. Therefore, in order to establish the impact that the introduction of European technologies and land use had on the environment, those studies that have compared long-term soil erosion records with land-use histories will be discussed here.

3.4.2 Colonial impacts in North America

Some of the first palaeoenvironmental studies undertaken on the impact of European settlement were from the USA, where European settlement first took place 200–300 years ago by French and English colonists. A summary paper by Webb and Webb (1988) reveals how the environment across eastern North America responded to the onset of

European settlement. The mean sediment accumulation rate for the period since European settlement from 129 sites from the mid-latitude region of the USA and Canada was 2.98 m ka^{-1}, nearly four times the rate recorded for the Holocene. Davis' (1976) early study of Frains Lake, Michigan, showed the impact of European settlement to be not only dramatic, but swift. Rates of erosion calculated from the lake sediment record showed an immediate increase in erosion from 0.1 to 5 t $ha^{-1}a^{-1}$ following forest clearance in the catchment. Importantly, her study demonstrated the initial vulnerability to erosion of the landscape during the transition to, and first years of, agriculture. In more arid regions of North America, the processes of deforestation and intense cropping were much more severe, leading to the infamous 'dust-bowls' of the western American prairies (Roberts, 1998).

Although the deforestation and cropping of land that followed European settlement in the USA are considered the main causes of the increase in soil erosion rates above those that characterised the pre-European environment, the introduction of large exotic mammals to these environments has also had a dramatic impact on land degradation. Trimble and Mendel (1995) summarised a large body of research on the effect of cattle on the environment in the USA, demonstrating their role as an 'important agent of geomorphological change'. High densities of cattle, or heavy grazing, are considered to cause the greatest impacts, reducing soil infiltration and soil vegetation, and increasing runoff and soil erosion. Unlike other large introduced mammals such as sheep, cows also frequently disturb the riparian zone, trampling river banks, and increasing erosion and turbulence (Trimble and Mendel, 1995).

3.4.3 Colonial impacts in South America

Research in South America has yet to generate the number of robust records (particularly from lakes) of the impact of European colonisation that now exist for the northern Americas. European colonisation in the region began earlier than nearly anywhere else in the New World (16th century). This has allowed records of European land use over much longer timescales to be captured, which may be of some significance in teasing out long-term environmental impacts. In the southeast of Brazil, European settlement in the 17th century began with wide-scale logging and cultivation of the land. Sediment cores from Jurujuba Sound record a distinct shift from re-worked marine sands deposited through most of the Holocene to terrestrial muds and sands, with sedimentation rates increasing sharply from this point onwards (Baptista Neto *et al.*, 1999). However, the greatest environmental impacts in this region occurred in the 20th century, with increased sedimentation rates and heavy metal contamination of the sediments. As with other New

World settlements, native vegetation was cleared and replacement by exotic species took place shortly after European contact.

In the northeastern pampas of Argentina, palynological studies demonstrate a significant decline in native grassland species coincident with the arrival of Europeans. Spanish *conquistadors* introduced a number of exotic herbivores and plants to the region in the late 16th century, which resulted in widespread vegetation disturbance, and increases in alluvial sedimentation rates, reflecting increased soil erosion (Prieto *et al.*, 2004). In Patagonia, European colonisation by *rancheros* began in the late 19th century, with Mancini (2002) identifying sheep and cattle grazing as the greatest cause of vegetation change during the entire Holocene period. This anthropogenic signal recorded greater environmental impacts than the combined effects of Holocene climatic fluctuations and *c.* 9000 years of pre-Hispanic settlement in the area.

3.4.4 Colonial impacts in New Zealand

In New Zealand, the impact of the first European settlers was preceded by that of the Maoris, who first arrived in mainland New Zealand from Polynesia after 1314 ± 12 AD (see also the discussion by Hogg *et al.* [2003]). Significant modification of the environment occurred from this time onwards. In particular, extensive burning of the landscape is considered to have greatly affected soils, and contributed to the rapid removal of forests from New Zealand (Wilmshurst *et al.*, 1997; Newnham *et al.*, 1998; McGlone and Wilmshurst, 1999). Page and Trustrum (1997) compared changes in sedimentation rates from Lakes Tutira and Waikopiro on the North Island of New Zealand to examine the erosional response of catchments to changes in land use. Sedimentation rates under pastoral land use (since European settlement) were 5–6 times greater than the rate under fern/scrub vegetation (*c.* 560 cal. BP to 1878 AD), and 8–17 times the rate recorded when the catchments were forested prior to the arrival of Polynesians (1850 cal. BP to *c.* 560 cal. BP). This suggests that the environmental response of the arrival of Europeans may have been muted by the fire-induced changes of the Polynesians hundreds of years earlier. The earlier sediments in the study are well dated by tephrochronology, but the identification of the point of European arrival in the sediments, based on the first appearance of exotic microfossil taxa and visual changes in stratigraphy, introduces some doubt as to the validity of the calculated post-colonisation sedimentation rates.

3.4.5 Summary

The arrival of Europeans in the New World took place in great many different environments, and over a long time period. The earliest colonisation occured in the 16th century, while Europeans were still arriving in near-pristine environments as late as the 19th and 20th centuries. Nevertheless, certain characteristics are common to all these experiences. The colonists brought with them exotic plants and animals that in many instances invaded and even replaced the native species. In the case of large herbivores such as cattle, these introduced species also directly altered the environment by increasing rates of soil erosion. Perhaps the most dramatic change was the implementation of Old World agricultural practices, which in most instances required extensive deforestation of native vegetation. Most depositional evidence of this period records increases in sedimentation, often by several orders of magnitude, in response to sharp rises in soil erosion. In some regions, these initial environmental disturbances were surpassed in magnitude during the 20th century, with mechanisation, increases in the intensity of agriculture and higher population densities (Baptista *et al.*, 1999), while in other locations the initial impacts of the Europeans instigated environmental shifts that continued for many decades afterwards (Gonzalez, 2001).

3.5 Developing long-term records of human-induced environmental change

3.5.1 Long-term records of catchment erosion in Australia

Numerous studies now exist that have estimated sediment yields from catchments across Australia in an attempt to assess the impact of human activity on erosion (e.g., Olley and Wasson [2003] and references therein). This work has typically focused on large-scale fluvial systems, utilising suspended sediment yield data and sediment budget analyses (e.g., Brookes and Brieley, 2004). Although such studies have provided coarse estimates of environmental change since European settlement in Australia in the late 18th century, the difficulties involved in estimating sediment flux by these means should be noted (even though they are rarely detailed in the literature). In most sediment yield studies, river systems have rarely been monitored for more than a few years. Suspended sediment yield data are also highly precipitation dependant (Duck and McManus, 1990), with not only seasonal but annual variability being exhibited in data sets. Furthermore, those measurements obtained are very unlikely to be representative of a catchment's sediment yield and transport processes over long timescales. In particular, they will almost certainly not be congruent with sediment yields from periods of different climatic or land-use characteristics.

The second general problem with the use of suspended sediment yield estimations is that of the inefficiency of the sediment transport process; most eroded soil material may potentially be stored within the catchment upstream of gauging stations (Walling, 1983). Estimates of suspended sediment yield also usually neglect the contribution of sediment from the catchment transported as bedload, and that being transported as dissolved load. Even in relatively small catchments, it has been shown that averaged sediment yields decrease with increasing catchment area, due to sediment storage (Neil and Mazari, 1993; Wasson *et al.*, 1998). At the scale of large fluvial systems, the inefficiency of the fluvial delivery system, and the extent of sediment storage in catchments makes calculating robust and meaningful estimates of sediment yield extremely difficult. As many authors have noted, little is known of where sediment moves within a basin, of how long it is stored in the landscape, of what volumes are retained, or why (e.g., Sutherland and Bryan, 1991; Beach, 1994).

Arguably the most critical aspect of sediment yield research is establishing the chronology of environmental change. Without a firm grasp of the timescales involved, the value of any information gained on rates of change or the timing of change is severely limited. The dynamic nature of the fluvial environment means that dating of sediments in these systems can be problematic; fluvial deposits very often experience re-working, and only episodic deposition. This is especially the case in Australia, where the behaviour of the fluvial system is closely linked to climatic fluctuations (e.g., flood- and drought-dominated regimes [Erskine and Warner, 1988]). While successful dating and reconstruction of fluvial sequences using radiocarbon methods have been achieved (e.g., Webb and Dragovich, 2004), the chronological resolution is typically insufficient to pinpoint exact historic events, such as the first arrival of Europeans in Australia. As such, the ability of fluvial sediments to provide detailed, high-resolution and complete sedimentary records of recent environmental change is limited.

Studies of closed lake basins that retain the erosional products of the surrounding catchment offer a solution to these problems. These basins are essentially sediment traps where a range of materials from the both the immediate catchment (e.g., soil, parent material, charcoal, pollen) and surrounding region (e.g., atmospheric pollutants, charcoal, pollen) may be retained over long time spans. By examining these sediments, a near-continuous record of environmental change may be reconstructed. Importantly, where such records predate European settlement in Australia, estimates of sediment yields under 'natural', or pre-disturbance conditions can be made (Dearing, 1991), providing an important benchmark against which subsequent environmental changes can be assessed.

3.5.2 Lake sediment records of environmental change in Australia

One of the earliest studies of lake sediments in Australia was undertaken by Kershaw *et al.* (1970) to provide information on past vegetation change. This has proved to be perhaps the most popular use of lake sediment records in Australia since then, with changes in the pollen content of sediments through time often being used to infer past climatic conditions. Other seminal studies in Australia have utilised lake sediment records to reconstruct fire histories, such as that of Singh *et al.* (1981) for the Late Cainozoic from Lake George in southeast New South Wales. Most relevant to this present research at Tocal Homestead Lagoon have been those Australian studies that have used lake sediments to provide estimates of past soil erosion, and lake records that have successfully archived the impacts of European settlement on the environment.

3.5.3 Lake and reservoir sediment records of soil erosion in Australia

The potential of lake sediments to preserve information on catchment soil erosion was first identified by Mackereth (1966), who concluded that 'One may regard the sedimentary sequence of a lake deposit as a series of samples of soils eroded from the drainage basin and deposited chronologically in the lake bed.' (Mackereth, 1966, p. 168). In Australia, recent decades have seen a proliferation of reservoir-based sediment research (e.g., Clark and Wasson, 1986; Moore, 1990; Erskine *et al.*, 2002). Much of this work has been undertaken on 'farm dams', usually small artificial ponds and lakes created on farms by damming streams, which have provided information on soil loss from agricultural land in Australia over decadal timescales. These studies have involved the measurement of volumes of sediment deposited in impoundments. Using the date of construction of the impoundment (when known), a single average rate of sediment accumulation is estimated for the period since the dam's construction. This value is then extrapolated across the area of the catchment to estimate rates of soil erosion (e.g., Neil and Fogarty, 1991). Estimating catchment-wide erosion using these procedures will generally deliver better results than studies based on the extrapolation of suspended sediment yield data alone, since the actual volume of sediment eroded from a catchment (assuming little is stored on the catchment's surface), and the time elapsed since erosion began are known. Very little of the work has been tested or evaluated for reproducibility, however.

As with measurements of soil loss using hillslope plots or fluvial suspended sediment yields, no long-term background or base level data are available for comparison when using reservoir sediments. Without such information the magnitude of soil erosion recorded after dam construction is nearly impossible to evaluate. In the case of impoundments constructed relatively recently (e.g., 20th century) (Erskine *et al.*, 2002), the processes of sediment transport and the erodibility of soils would most certainly have been modified prior to dam construction. The sedimentary record from such sites will provide information only on catchment behaviour following disturbance.

3.5.4 Sediment yield reconstructions

Moore's (1990) and Caitcheon's (1990) studies of the sedimentary infill of two 'farm dams' in northern South Australia and southern New South Wales were of great importance in providing some of the first data on the impact of land use on catchment soil erosion in those regions. Unfortunately, neither dam was constructed until the early 20th century, allowing only relatively recent land use impacts to be assessed. As such, their records fail to record either the longer-term patterns of soil erosion, perhaps driven by climatic processes, or the initial impacts of agriculture on erosion.

Neil and Fogarty's (1991) survey of the sedimentary infill of 46 farm dams on the Southern Tablelands of New South Wales improved upon previous work by providing some measures of reproducibility. Sedimentary infill in the dams was very recent, however, with the dams surveyed ranging in age from only three years old to a maximum of 40 years. It must be noted that such short-term measurements of sediment yield provide no information on the catchment's behaviour over long time spans. Some measure of the impact of modern land-use on erosion was included in the study. 'Natural sediment yield rates' were estimated using the sediment yields calculated from four dams in undisturbed catchments as a proxy record of the behaviour of the catchments prior to European settlement. Although methodologically this presents an improvement upon taking no account of pre-disturbance environmental conditions and soil loss, these sites cannot account for differences in Aboriginal land use, fire regimes, changes in vegetation, and differences in soil properties that characterised the pre-European environment. Nor can this approach account for changes in climatic conditions since the European settlement (e.g., Tibby, 2003). Where records of sediment yield exist that cover the earliest years of environmental modification following first European settlement, it has become obvious that the most severe erosion and degradation took place in the first several decades of contact, and that rates of 20th century change are in contrast quite low (e.g., Wasson and Galloway, 1986). It is therefore essential that for meaningful quantifications of rates of sediment yield to be made, both recent and long-term records of sediment yield be available at a single site (Dearing, 1991). Unfortunately, research into long-term, continuous lake records of sediment yield has been rarely undertaken in the Australian environment and few studies have been published that accurately estimate catchment sediment yield through the late Holocene.

3.6 Records of the impact of European settlement in Australia

3.6.1 Sedimentary records of European impacts in Australia

Dodson and Mooney (2002), in a paper titled 'An assessment of historic human impact on south-eastern Australian environmental systems, using late Holocene rates of environmental change', have presented a summary of sediment records that contain information on environmental change in the late Holocene. Although all these studies record changes in sedimentation rates and pollen assemblages in the last *c.* 500 years, only one of them (Haworth *et al.*, 1999) has a reliable and detailed chronology for the period of European settlement that is sufficient to provide estimates of sedimentation rates. Even then, Haworth *et al.*'s (1999) record from Black Mountain Lagoon on the New England Tablelands of northeast New South Wales provides sedimentation rates only from the early 20th century onwards, missing the earliest years of colonial landscape modification, and quite possibly the greatest impacts on the environment to date (Gale and Haworth, 2005).

The data presented by Dodson and Mooney (2002) represent a common failing of studies of European impacts in the Australian environment. Most late Holocene sedimentary records from Australia have lacked a chronology of sufficient resolution to allow interpretations of specific events on the environment to be made. Because of the indistinct timing of changes in these sedimentary records, shifts in environmental proxies assumed to be the result of European settlement could equally be the result of Aboriginal activity, or sub-millennial climatic variations during the last 500 years. At worse, changes in these records thought to be coincident with European settlement may be artefacts of disturbed or discontinuous records; several of the sites in Dodson and Mooney's (2002) compilation have evidence of reworking of the sedimentary records (e.g., Johnson, 2000). Unfortunately, in most of these cases there is no way of knowing whether changes interpreted as the record of European settlement are the result of this; most of these studies would be much better utilised as sources of mid- to late Holocene environmental changes alone, as their chronologies for this longer timescale (usually provided by ^{14}C dating) are more reliable.

The presence of the pollen of the exotic tree *Pinus* has been thought to accompany first European settlement in Australian sedimentary records (e.g., Mooney *et al.*, 1997). However, as Tibby (2003) has shown, a range of studies now demonstrates that a significant lag time may exist between the first introduction of *Pinus* in the landscape (as revealed in documentary sources) and its first appearance in the sedimentary record.

Though Mooney and Dodson (2001) presented estimates of the impact of first European settlement in southwestern Victoria, their chronology is questionable since it is based solely on the first appearance of *Pinus* pollen in the sediments of Lake Keilambete. Once deposited, pollen may also be reworked, and displacement in the vertical plane has been reported. Dodson *et al.* (1993), for example, recovered *Pinus* pollen in lake sediments dated to 600 cal. BP, half a millennium before European settlement in the region. Although this is probably due to an unreliable ^{14}C date, the reworking of pollen downcore cannot be ruled out. This relatively new finding is of enormous significance in Australian sediment-based studies of human impact, and unfortunately casts some doubt on the reliability of findings where the first appearance of exotic pollen is the primary or sole source of chronological information (e.g., Mooney and Dodson, 2001).

A problem common to many of those studies that have attempted to develop records of the impact of European settlement on the Australian environment is that of site selection. Many studies included in Dodson and Mooney's (2002) review paper seem to record very low rates of sedimentation for the late Holocene, and more importantly for the period since European arrival. As a result, the resolution of these environmental records may often be insufficient to record human impacts. In Mooney *et al.*'s (2001) study of Jibbon Lagoon in Sydney, New South Wales, the sediment estimated to have been deposited since European settlement is just 19 cm thick. A possible solution to this problem may involve very high-resolution sampling, such as that employed by Harle *et al.* (2002) in Lake Dora in western Tasmania. A core from this low-sedimentation lake was sub-sampled at intervals as fine as 2.5 mm. Selecting study sites that display evidence of high sedimentation rates, however, offers researchers not only a high-resolution record, but the added benefit of sufficient sediment to allow a range of proxies (pollen, chemistry, magnetic properties etc.) to be measured. Stanley and De Deckker's (2002) Holocene record from Blue Lake, in the Snowy Mountains of southeast Australia, for example, is characterised by high sedimentation rates, so that 1 cm thick samples provide palaeoenvironmental data at approximately 25 year intervals.

3.6.2 Well-dated sediment records of the impact of European settlement

Lead-210 dating of the Lake Dora core by Harle *et al.* (2002) provided a record of vegetation change since sometime before 1846, and perhaps as early as 1811. The ambiguity of the lowest sediment ages is a consequence of dating the basal part of the sequence by extrapolating the overlying sedimentation rate downwards. Assuming a constant sedimentation rate for this period, sediments from 7.5 cm depth would have been deposited in *c.* 1811. Surprisingly, the results from Lake Dora show very little variation in sediment properties, sedimentation rates or pollen taxa during the period of first European settlement, though some shifts in these proxy indicators were recorded later in the 19th century. This study offers support for those Australian researchers who have proposed that the initial impacts of European settlement on the Australian environment were essentially minor compared with those changes instigated later in the 19th and 20th centuries (Clark, 1990). However, a large body of historical and palaeoenvironmental evidence suggests that the arrival of Europeans in Australia had profound and rapid impacts on the environment.

A broad-scale (both chronologically and spatially) picture of the impact of European settlement on vegetation change may be gained from the compilation of late Holocene pollen records from 71 sites in southeast Australia by Kershaw *et al.* (1994). The main finding from these data is that after European settlement the incidence of *Casuarina* typically decreased, while grass pollen increased sharply. This suggests an increase in grassland across the region, probably in response to deforestation efforts by the colonists. As Gale (2003) has shown, this broad change in vegetation echoes much of the historical evidence of vegetation change from the early colonial period.

Gale *et al.* (1995) used ^{210}Pb methods to accurately date lake sediments from Little Llangothlin Lagoon on the New England Tablelands of New South Wales, locating the depth at which European settlement was recorded in the lake. In a later paper, Gale and Pisanu (2001) presented a vegetation history for the late Holocene period from this site, and reconstructed the changes in vegetation that followed European settlement. The pre-European pollen assemblage was dominated by Casuarinaceae, which declined dramatically following European settlement with Mimosaceae and Myrtaceae replacing the Casuarinaceae. The impact of the Europeans on vegetation here was swift; these changes took place within the first 25 years of European settlement. The work of Gale and Haworth (2005) discussed the record of catchment-wide soil loss for the pre- and post-contact period obtained from the pattern of lake sedimentation in Little Llangothlin Lagoon. Their work provided evidence of the ferocity of the initial impacts of European settlement on soil loss, which in this catchment led to sediment starvation in the surrounding catchment, and sediment supply limitation of erosion rates within decades of their arrival. Providing quantitative evidence of European impacts, they found that rates of catchment-wide soil loss increased by over 50 times immediately following the arrival of the first sheep with European settlers in the catchment.

Gale and Haworth (2002), and subsequently Gale *et al.* (2004), demonstrated the importance of high-resolution, well-dated and continuous records that span the period of first European settlement in Australia. Using a ^{210}Pb dated record that included several decades of sedimentation before the first official date of settlement in 1837, Gale *et al.* (2004) showed that significant environmental disturbance took place in the catchment of Little Llangothlin Lagoon immediately before this date. Gale and Haworth (2002) suggested that this was the result of either Aboriginal activities or, more plausibly, that the 'shadow of Europeans' had fallen onto the site well before the 'official' date of settlement. Gale *et al.* (2004) provided additional evidence to support the latter explanation. These results show that the timing and nature of European settlement in Australia and its impacts may be much more complex than commentators have previously thought.

There is some evidence that the impacts of European settlement may have been even more severe in the arid zone of Australia. Wasson and Galloway (1986) calculated a long-term, pre-European sedimentation rate for Umberumberka Creek in western New South Wales (Figure 3.2), which they then compared to a post-settlement rate determined from a reservoir on the same creek, built during the colonial period. Sediment yields increased by 66 times during the post-settlement period, from a mean rate of 0.037 to 2.46 $m^3 ha^{-1} a^{-1}$ (Table 3.1). Figure 3.6 shows the changes in sedimentation for the last 1000 years at Little Llangothlin Lagoon, Redhead Lagoon and Umberumberka Creek demonstrating how dramatic the increase in sediment yield has been in the post-settlement period in eastern Australia. More importantly, these studies show that the greatest rates of erosion to date accompanied the first decades of European settlement in the 19th century (Figure 3.6).

Table 3.1 Increases in sedimentation rates following European settlement in eastern Australia. Sedimentation rates calculated by Gale (2003).

Site	Mean prehistoric sedimentation rate	Mean post-settlement sedimentation rate	Increase in sedimentation rates	Source of data
Little Llangothlin Lagoon, northeast NSW	70 $t\ km^{-2} a^{-1}$	474 $t\ km^{-2} a^{-1}$	**11**	Gale and Haworth (2005)
Redhead Lagoon, central eastern NSW	0.012 $kg\ m^{-2} a^{-1}$	0.51 $kg\ m^{-2} a^{-1}$	**43**	Williams (2005)
Umberumberka Creek, western NSW	0.037 $m^3 ha^{-1} a^{-1}$	2.46 $m^3 ha^{-1} a^{-1}$	**66**	Wasson and Galloway (1986)

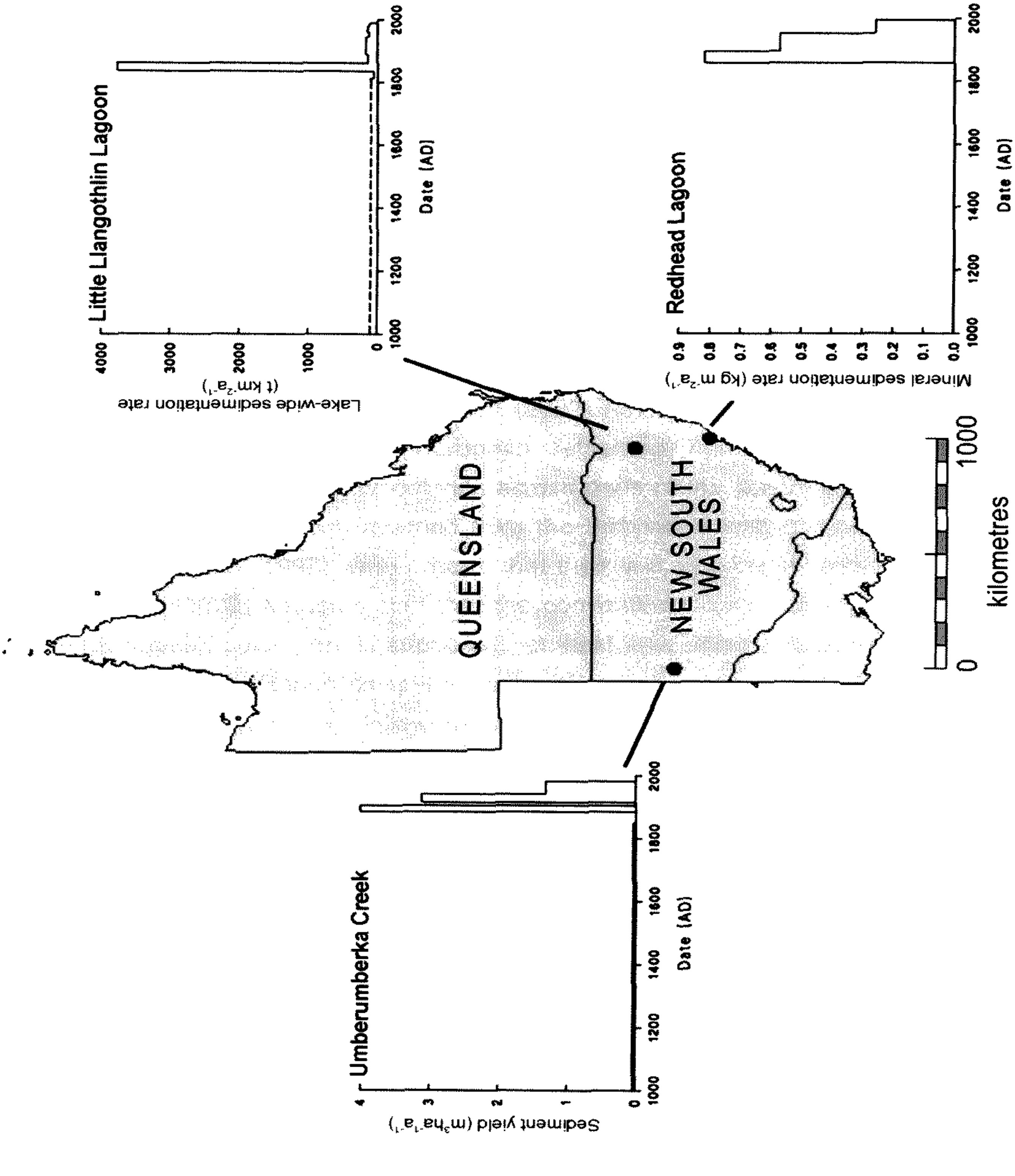


Figure 3.6 Sedimentation rates at Umberumberka Creek, Little Llangothlin Lagoon and Redhead Lagoon, New South Wales for the last 1000 years. Adapted from Gale (2003).

3.6.3 The historical record of European impact on the environment

The observations of early explorers in Australia shortly after the first wave of European settlement have been compiled by a range of authors in an attempt to characterise the impact of Europeans on the environment. Historical sources have been used in particular to examine how and when vegetation was modified, how soils were modified and what impacts European land use had on erosion. Two main schools of thought may be identified in the literature concerning the outcomes of European settlement on vegetation (Gale, 2003). Earlier work by authors such as Adamson and Fox (1982) proposed that both native grasses and forest species were cleared to make way for large-scale pastoralism. Forest species were then progressively replaced by grasses, with introduced species becoming dominant by the 20th century. Rolls (1999), Flannery (1994), Ryan *et al.* (*c.* 1995) and Noble and Grice (2002) have argued that, in contrast, major changes in fire regime following European settlement led to a reduction in native grasslands and an increase in woodlands across parts of the continent. This latter view, based mainly on observations quarried from the archival record, has been attacked by Benson and Redpath (1997), who argued that their use of such observations has been selective. Gale (2003) demonstrated that the conventional theory of rapid deforestation following European settlement is supported, at least in southeast Australia, by the fossil pollen record and by other compilations of historical observations. Similar shifts in vegetation are thought to have followed European settlement in the Hunter Valley region (Bennett and Mooney, 2003).

Compilations of historical reports on the state of soils in Australia shortly after European settlement have been used to provide evidence of European impact on soil erosion (Gale, 2003). The overwhelming picture is that pre-European soils are likely to have been 'spongy', uncompacted, fine-grained and often organic-rich. The displacement of Australian native species by hard-hoofed domesticated species resulted in soil compaction and reduced infiltration capacities, with subsequent increases in runoff and erosion (Gale, 2003). The impacts of these changes in the landscape quickly became visible to the colonists, who documented the appearance of gullies and bemoaned the rapid silting up of rivers, which were of enormous importance for transport in the early colony. Early reports of environmental degradation in the Hunter Valley region are discussed later in this thesis (section 4.4.5).

3.7 Summary

The Holocene period in Australia has been characterised by a range of climatic fluctuations and human impacts; the precise timing, magnitude and location of which are clearly not entirely understood at present. Furthermore, it would be fair to say that in the late Holocene, and particularly the last 1000 years, the detail of information about past environmental conditions becomes increasingly poor, undoubtedly due to the lack of high-resolution records from this crucial period of environmental history. In the past, studies, particularly of Quaternary-scale environmental shifts, have perceived the Holocene in Australia to have been a period of relatively minor changes. However, with an increased understanding of the impact that European settlement had on the environment, some researchers have challenged this view, arguing that the environmental shock of the 18th century was in some respects (mainly relating to vegetation and soil impacts) equal to those changes experienced during the glacial maximum in Australia.

Another school of thought argues that the initial impacts of European settlement on ecosystems were relatively benign, and that more significant changes occurred much later (in the late 19th or 20th centuries). Evidence from studies in other New World locations lends support to both arguments to some extent, but the overwhelming experience worldwide appears to be that the arrival of Europeans and the onset of European land use brought about rapid and major environmental changes. It may be that the particular pattern of European settlement in Australia led to a history of environmental change, with more minor initial impacts, that is unique in the New World. The historical record has been of some use here in providing information for the debate, but much of the archival resources are spatially limited and of marginal relevance. Clearly, in Australia a much greater range of sites across the continent needs to be investigated before an adequate understanding of the matter is gained.

Chapter Four

Land use history

4.1 Introduction

The catchment of Tocal Homestead Lagoon has been under European land-use from the early 19th century to the present day. A detailed record of the environmental history of the catchment and the immediately surrounding areas has been constructed from an array of sources, including explorer's journals, newspaper articles, historical stock and cropping records, and visual media including aerial photographs, amateur photographs and paintings. Much of this work has also been presented by Cook *et al.* (in press). As the Tocal estate has been in the ownership of only three families since the property was first taken up in 1822, followed by management by government authorities, detailed records have been preserved. The general aim of reconstructing the environmental history of Tocal has been to identify the timing, location and scale of anthropogenic activities that may have altered the surrounding environment. The environmental history will attempt to characterise the conditions that prevailed prior to, and following European contact while also focusing upon the impact of the first several decades of agriculture at the site. Obtaining and evaluating what little information exists on the condition of the environment at this early stage of colonisation is of vital importance if any comparison with the subsequent decades of environmental change are to be made.

The specific aims of this chapter are to:

(1) Identify when Europeans first entered, and subsequently settled in the Tocal catchment

(2) Synthesise the limited amount of published archaeological data for the surrounding area, and to reconstruct the pre-European history of Tocal

(3) Document how land use has changed within the catchment during the last two centuries, and obtain quantitative data, such as stocking rates and human population densities, that represent these changes

4.2 Pre-European human activity in the Hunter valley

4.2.1 Introduction

Although the size of Australia's Aboriginal population during the Holocene is the subject of a great deal of debate (e.g., Webb, 1984), it is likely that population densities were generally higher than those detailed in the historical record for the period immediately following colonisation. Historical records of the Aboriginal population (such as the 1828 Census of New South Wales) are imperfect indicators of pre-colonial population densities as the devastating effects of the smallpox virus (such as that experienced by the Port Jackson Aborigines in 1789 that is thought to have originated in northern Australia as a result of contact between Aborigines and Macassan traders [Campbell, 2002]) and conflict with the Colonists (Rowley, 1970; Lines, 1991) (but see also Windschuttle, 2002), as the importance of European-Aboriginal conflict in reducing Aboriginal numbers is currently keenly debated) may already have significantly reduced the population before any data were collected. European-Aboriginal conflict, particularly in the upper Hunter valley was not uncommon (Anon., 1824; Anon., 1826; Wood, 1972, p. 130).

The idea that populations were probably much higher during the late Holocene is supported by radiocarbon dates from archaeological sites across Australia (Lourandos and David [2002] and references therein). However, as Lourandos and David (2002) have noted, the central Queensland highlands and southeast Queensland show a marked downturn in numbers of dated Aboriginal sites from around 1000 years ago that may predate, and therefore be independent of, the impact of European colonisation.

4.2.2 Archaeological evidence

Several locations in the Hunter valley provide evidence of increasing Aboriginal numbers during the late Holocene. The number of occupied sites in the Upper Mangrove Creek catchment, located southwest of the Hunter valley, increased during the mid- to late Holocene (Attenbrow, 1987). At several sites in the catchment (including Mussel Shelter) increasing artefact accumulation rates suggest that human activity increased between 3000 cal. BP and 1500 cal. BP than during earlier or later periods (Hiscock, 1986). The oldest date of human occupation in the Hunter region is from Loggers rock shelter in the catchment of Upper Mangrove Creek (*c.* 11 000 cal. BP [Attenbrow, 1982]). In the semi-arid region inland of the Hunter valley, the maximum age of human occupation on the Liverpool Plains is estimated as late Pleistocene (*c.* 19 000 cal. BP) by Gorecki *et al.* (1984), while at Cuddie Springs in central northern New South Wales, first human occupation is thought to date from around 30 000 years ago (Field and Dodson, 1999). A better geographical relationship is gained from the data set compiled by Lourandos and David (2002) for the greater Sydney region and those few dated sites from the Hunter valley. Of 82 radiocarbon dates from 27 sites, no sites have been recorded for the last glacial maximum or earlier. The number of dated sites increases during the terminal Pleistocene, with a small peak at *c.* 7000 cal. BP, while the greatest numbers of sites date from *c.* 2000 cal. BP (Lourandos and David, 2002). This trend is similar to that recorded for the temperate regions of southeast Australia, suggesting that the pattern of population trends in the Hunter valley may have been similar to those of the temperate southeast, rather than those of the arid zone (inland of the Liverpool Ranges [Gorecki *et al.*, 1984]). As most Hunter valley sites have been occupied for much less than 10 000 years, the region appears to support the Holocene Aboriginal 'intensification' model of Lourandos (1983). Some mention needs to be made of the findings of Nanson *et al.* (1987), who presented evidence of Aboriginal activity in the Nepean River valley, south of the Hunter valley, from over 40 000 years ago. Recent dating of an artefact from the site extends the proposed maximum age of activity to nearly 50 000 years ago (Stockton and Nanson, 2004). As there are no major geomorphic boundaries or obstacles between the two regions (quite the opposite in fact, as the first European overland route from Sydney to the Hunter valley followed a well established Aboriginal trade route between the regions [sections 4.3.2 and 4.3.3]), it is curious that no evidence of Aboriginal activity in the Hunter valley has been found from earlier periods.

The Hunter valley region is generally considered to have been populated by three Aboriginal tribes. The upper Hunter was the territory of the Geawegal people, the mid- to lower Hunter valley was occupied by the Wonarua people, while the estuarine Hunter River area is thought to have been inhabited by the Gaddhang or Worimi tribe. Early writers on the Aboriginal populations of the Hunter valley region have estimated the number of Wonarua people at the time of first contact to be between 500 and 600 (Moore [1970] and references therein).

4.2.3 Early observations of Aboriginal populations

Little is known of the history of Aboriginal occupation in the immediate region of Tocal. Aboriginal tribes are said to have inhabited the Tocal area for 'probably ... 7000 years or more' (Archer, 1988). In truth, no evidence has been presented to date that can address the question of how long, or in what numbers Aboriginal people had occupied the immediate site. The annotations of Captain James Cook from 1770 represent the first observation we have of Aboriginal occupation in the Hunter valley region. At nearby Port Stephens (Cook, 1955, p. 314), he recorded the following:

> We saw several smooks a little way in the Country rise up from the flat land, by this I did suppose that there were Lagoons which afforded subsistence for the natives ...

Many such observations were made by Cook as he sailed north along the east coast of Australia in 1770, and most relate to cooking fires. However, Benson and Redpath (1997) are incorrect in their assertion that Cook observed only cooking fires on this voyage. Importantly, Cook's diary records an incident near the Endeavour River in far north Queensland when Guugu Yimidhirr people purposely started a large fire in grasslands following a disagreement with the Europeans over a turtle (Thomas, 2003). The observations of Lieutenant-Colonel William Paterson's expedition in the lower Hunter valley in June and July 1801 represent the earliest data from the post-settlement period that we have on Aboriginal occupation. Based on the large numbers of canoes he saw, expedition member Francis Barrallier concluded that there were 'great numbers' of Aborigines in the area (Barrallier to Greville, 1801).

Unfortunately, the November 1828 Census of New South Wales is the only known survey of Aboriginal population numbers in the immediate area from the early colonial period. The 1828 census records the Aboriginal population of the nearby Mt Johnson and Old Settler's areas as 53 and 51 respectively. One source listed the Aboriginal population of the lower Paterson valley following settlement to be 132 (White, 1986, p. 141). The number of Aboriginals recorded attending the annual issue of blankets at Paterson (Threlkeld [1834; 1838-1840], cited in Brayshaw [1986]) from 1834–1840 (Table 4.1) suggest a population for the Paterson Plains area of less than 100 during this decade. Population numbers declined rapidly after this time to perhaps 500–600 over the entire 5000 km^2 of the upper Hunter valley by the 1840s (Miller, 1887, p. 352; Fawcett, 1898, p. 152). By 1854, the *Maitland Mercury* reported that 'The last of the Newcastle Tribe' had died (Anon., 1854).

Table 4.1 The number of Aborigines who attended the Paterson blanket distribution centre, New South Wales, 1834–1840.

Year	**Adults**		**Children**	
	Male	*Female*	*Male*	*Female*
1834	18	10	2	
1838	29	20	6	2
1839	27	11	3	1
1840	39	4	6	

Several researchers have suggested that water bodies have been foci of Aboriginal activity in the Australian environment, especially during periods of drought (Luebbers, 1975; Grothen, 1988; Hiscock, 2000; Gale and Haworth, 2002). Several examples of this relationship in the lower Hunter valley were recorded by the area's first explorers. While visiting the Hunter River at nearby Maitland and Hexham in 1827, Peter Cunningham described in great detail many lagoons located on the river's floodplain. He also noted the concentration of Aborigines around the floodplain swamps and lakes (Cunningham, 1827, pp. 150–151):

> These lagoons swarm with the most delicate fish; and during the dry summers, when the water is low, the natives wade in and actually drag out *cart-loads* thereof ...

Colonial artist Joseph Lycett (1824) made similar observations in the Hunter valley. McKieran (1911, p. 890) detailed the recovery of Aboriginal weapons by workmen digging a drain through a swamp in nearby Raymond Terrace. At Hexham Swamp, a large floodplain swamp connected to the Hunter River, the Pambalong Tribe was recorded as living around the water body (Maynard, 2000), and Grothen (1998) has documented the substantial Aboriginal occupational history of Glenrock Lagoon, a flood-plain lake on the Hunter River near Maitland. Since Tocal Homestead Lagoon is a substantial permanent water body in close proximity to these sites this would suggest that it too was probably a site of Aboriginal activity.

4.2.4 Evidence of pre-European occupation at Tocal

Very little documentary record exists of an Aboriginal population located within the catchment of Tocal Homestead Lagoon. Most historians imply that the area surrounding the lake was occupied by Aboriginal people prior to the arrival of Europeans (Wood, 1978; White, 1986; Archer, 1988; Hunter, 1997), most probably by the Gringai clan of the Wonnarua people (Miller [1985] cited in Walsh [1999]). Although no prehistoric archaeological survey of the property has been published to date, several authors have cited a series of grinding grooves preserved in an outcrop of sandstone approximately 500 m west of the lake as evidence of former Aboriginal occupation (e.g., Walsh, 1999) (Figure 4.1). The proximity of the grooves to the lake is typical according to Brayshaw (1986, p. 94), as water is needed during the axe grinding process. Two other sites with grinding grooves have been found at Tocal, and a third on the *Stradbroke* property on the opposite side of the Paterson River from Tocal (Laffan and Archer, 2004). More extensive grinding grooves ('... one hundred and sixty places', Enright [1936]) have also been found at Greenwattle Creek, a tributary of the Paterson several kilometres east of Tocal (Figure 4.2).

McKiernan (1911, p. 890) detailed the use of stone tools by the Aborigines of the lower Hunter River valley, stating that they possessed '... stone hatchets, which were sharpened by other stones ...' and that '... large axes were beautifully ground and polished and an amount of trouble was taken in fixing handles to them. Years were often spent in grinding them ...'

Figure 4.1 Axe-grinding grooves in a sandstone outcrop, in the Tocal Homestead Lagoon catchment, 1998.

Figure 4.2 Axe-grinding grooves, Greenwattle Creek, Paterson. (Source: Newcastle Public Library).

Further evidence of probable pre-colonial Aboriginal activity at Tocal comes from the discovery of stone tools around the lake's perimeter (Archer, A.C., Principal of Tocal Agricultural College, personal communication, 1999) and from Aboriginal artefacts uncovered during excavations on the lagoon's slopes (Laffan and Archer, 2004). Such discoveries are not uncommon in the region: from the early days of colonisation in the lower Hunter valley settlers had, on occasion, found edge-ground axe-heads when ploughing their farms (Moore, 1970). In a report in the *Sydney Gazette*, geologist W.J. Enright (1923) described the recovery of a stone axe, found *in situ* at a depth of '11 feet from the surface' in West Maitland, securing the axe's antiquity as pre-European. A final clue to the Aboriginal history of the site may be inferred from the site's name of Tocal, thought to be derived from the Aboriginal word *Tugal*, which according to King Jacky Jacky of the Pandago tribe meant 'big' or 'plenty' (Boydell [1835] cited in Walsh [1999]). Several authors have suggested that this epithet was a reference to the lagoon itself (e.g., White [1986]).

4.2.5 Aborigines and fire in the lower Hunter valley

Early accounts by missionaries and settlers in the Hunter valley record the widespread use of fire by the local Aborigines. Sokoloff (1978a) wrote that the Aborigines of the Hunter valley used fire for signalling, ceremonial purposes, fishing (see also Palmer [1893] and Sokoloff [1978b]), hunting and vegetation modification. The last of these, the so-called 'fire-stick farming' of the landscape by Aborigines, has been the subject of considerable debate (popularised by writers such as Flannery [1994]). Although Sokoloff's writings on fire in the area are not contemporaneous with the events, the following observation of firing of the landscape by the Hunter valley Aborigines, recorded by Parry (1831, p. 61), is:

> I never saw anything like the state of the country with the fires – literally as black as charcoal for miles together. I am confident we rode four of five miles at once without seeing 100 acres of grass.

4.3 European settlement in the Hunter valley

4.3.1 Early and unofficial European occupation

The first Europeans to enter the Hunter valley region were escaped convicts from the Sydney penal colony, who did so before the end of the 18th century. White (1986) discussed the first European contact in the region, importantly noting the possible exchange of technology from the Europeans to the Aborigines:

> Until 1801 there had been few white men to disturb the Aborigine's peace, only some Europeans who having been shipwrecked sought refuge by exchanging tools for hospitality.

This brief account probably refers to the five convict escapees from the Sydney penal colony who landed in nearby Port Stephens in 1790 (Collins, 1798). They are considered to be the first Europeans to arrive in the Hunter valley (Cook *et al.*, in press). The convicts are recorded as having lived with local Aboriginal people until 1795 when they were picked up by the ship 'Providence' (Collins, 1798, pp. 425–426). A similar, but entirely separate record of early pre-official European contact is that of nine convicts and two children landing at Swansea and then either Newcastle or Port Stephens in 1791 (Martin, 1991). The writer(s) of 'Memorandoms', attributed to a James Martin, detailed how a group of 11 escapees from Sydney Cove landed and made contact with natives at several locations on the coast of the Hunter valley region. During this contact 'clothes and other articles' were given to the local inhabitants. It is thus possible that the transfer of European technologies to the Hunter valley Aborigines took place prior to the official settlement of the area in the early 1820s (section 4.3.2). Since such exchange did occur, then the extent of vegetation and catchment modification in the decades preceding the first European settlement in the region may be greater than would be otherwise expected.

4.3.2 European exploration and descriptions of the lower Hunter valley

The first official exploration of the lower Hunter was by Lieutenant-Colonel William Paterson in June and July 1801. Within the limited time they had available, his party ventured about 20 km upstream of the present town of Maitland (Barrallier to Greville, 1801; Barrallier to King, 1801; Grant, 1801; 1803, pp. 149–166; Harris to King, 1801; Paterson, 1801; Paterson to King, 1801). Paterson reported the sighting of a person 'believed to be a European' (Paterson to King, 1801, p. 416) and, from evidence of blade marks on trees, he believed that 'some of the European deserters' were in the area (Paterson, 1801, p. 178).

However, it is equally likely that the marks were the work of Aborigines, for Francis Barrallier, recorded that 'I found in their Kenou Little hand hatchets English [in] manufacture' (Barrallier to Greville, 1801, p. 82). Notwithstanding the evidence of the hatchets, Barrallier believed that 'Europeans have frequented this place only 5 or 6 in Number' (Barrallier to Greville, 1801, p. 82).

The dispersion of European technologies in eastern Australia ahead of the Europeans themselves has been previously discussed by Gale and Haworth (2002, p. 133), as have the implications of this for environmental disturbance. For example, Dawson (1830, p. 135, cited in Gale and Haworth [2002]), recorded observing an iron tomahawk in the possession of Aborigines north of present day Port Stephens in 1826. In the Hunter valley, European technologies, such as those English-manufactured axes observed in June 1801 by Barrallier are thought to have arrived along Aboriginal trade routes from the Sydney basin (McCarthy, 1939; Moore, 1981), or perhaps by European escapees as previously detailed.

Barrallier returned to the Hunter in November 1801 with the Acting Surveyor-General, Charles Grimes, to complete the survey of their discoveries, this time exploring some distance up the Paterson River (Grimes, 1801). In the light of the enormous environmental impact that must have occurred over the next few decades as a result of logging riparian red cedar (*Toona ciliata*) along the Hunter, it is it is pertinent to note Grimes' (1801, p. 415) observation that the lower Paterson possessed more cedar than any other part of the river system. Peter Cunningham, describing the condition on an unsettled stretch of the lower Hunter, recorded that beyond the densely wooded riparian zone, on the alluvial plains of the lower Hunter valley 'there is scarcely a superfluous tree to be seen, not often above a dozen per acre.' (Cunningham, 1827, p. 156). Also, that 'The alluvial banks of the Patterson's (*sic*) and William's Rivers are heavily timbered, but the forest land behind is open, grassy, and every way suitable for pasture without cutting down a single tree' (Cunningham, 1827, p. 147).

4.3.3 Early European settlement in the Paterson region

The earliest account of colonial habitation on the Paterson River near Tocal is of '6 well behaved convicts and two free men' who had established farms on the river banks by 1813 (Macquarie, 1956; Hunter, 1997). Governor Macquarie, shortly after his visit to the Newcastle penal settlement, initiated the first official access to lands in the Hunter valley for logging and limited farming (White, 1986). Convict teams from Newcastle had been logging the riparian red cedar for approximately 10 years prior to this, typically along the Hunter between Newcastle and Wallis Plains/Maitland (Archer, 1986; Wood, 1972). By July 1818, there was a cedar camp near the junction of the Hunter and Paterson Rivers (Macquarie, 1956, p. 132) and by the same year, a convict cedar camp known as Old Banks had been established on the banks of the Paterson (Archer, 1986, p. 3; Hunter, 1997, p. 13). This was located just downstream of Tocal and was presumably the base for logging operations all along this part of the river.

Meanwhile, escapes from the penal settlement at Newcastle were becoming more common. Many escapees were re-captured, some were probably killed by Aborigines, some probably died in the remote and difficult country that surrounded the settlement and some successfully made their way back to the Hawkesbury (Macquarie to Bathurst, 1819a, p. 43; Morisset, 1820, p. 462). It is likely, though, that several survived, either with the Aborigines or in gangs with other escapees. If this is the case, were their numbers sufficient to cause environmental disturbance that may be identified in the palaeoenvironmental record (see Gale and Haworth [2002] and Gale *et al.* [2004] for a discussion of this question from elsewhere in Australia)?

Probably in 1812 and certainly by May 1813, four well-behaved convicts were permitted to occupy land at Paterson's Plains (Perry, 1963, p. 61). By July 1818, there were eight farms, two occupied by free men and six by convicts (Macquarie, 1956, pp. 130–131). By 1820 there were 12 farms (Morisset, 1820, p. 477). As early as 1818, Governor Macquarie (1956, p. 131), on a tour to the then remotest part of his bailiwick, noted that there was '... much more ground cleared & cultivated than I had any idea of.'

Joseph Lycett accompanied government officials to Newcastle and the Paterson River in 1819, where he painted and recorded his observations of the landscape, flora and fauna and the indigenous population (Figure 4.3).

Figure 4.3 Reproduction of Joseph Lycett's 'Lake Patterson, near Patterson's Plains, Hunter's River', a large lake on the eastern bank of the Paterson River, in the vicinity of Tocal Homestead Lagoon, from Turner (1997).

Although his prints and descriptions have been considered a useful source of information on the natural environment of colonial New South Wales (see, for example, Archer [1986]), he had been convicted of forgery and was a liar, with artworks attributed to him discovered to be the work of others (Turner, 1997). The illustration by Lycett in Figure 4.3 of Lake Patterson is of a water body that is thought to have since been drained (Archer, 1986); the notes that accompanied the engraving are as follows:

> This tract of land is one of the finest in this vast country. It contains many thousand acres of the most beautiful grass, well watered, and is, in general, thinly wooded. The land … near the sides of the river, is particularly good, and produces amazing crops of wheat, maize, barley, oats, and potatoes of the best and largest sort, with every kind of vegetable …' (Lycett, 1824)

4.3.4 The first surveys and early occupation of the Tocal catchment

The government surveyor, Henry Dangar, surveyed the Hunter valley area in 1822–1823, producing the first documented map of the region in 1823. Historical records and a recently discovered unpublished map of the middle Paterson River area (attributed to Surveyor O'Brian, and dated between 1820 and 1826 [Hunter, 1997]) (Figure 4.4) suggest that not only was the riparian zone at Tocal logged prior to Dangar's 1823 survey, but that farming may already have been under way within the catchment of Tocal Homestead Lagoon. The 1820s saw an increasing stream of free settlers and arriving in the Hunter valley to take up property for farming. Amongst these was George Pell, who according to Hunter (1997, p. 43), may have been the occupant of the dwelling located in the catchment of Tocal on the pre-1823 map (Figure 4.4). Dangar's survey also records an unnamed property on the same bank of the Paterson, which had 15 acres cleared '… by a settler who has since left'. Hunter (1997, p. 43) has postulated that the abandoned farm recorded by Dangar is that depicted in the pre-1823 map, on the present day site of Tocal. This would be the first recorded European occupation of that site.

Early farming in the southern portion of the Tocal site may have been undertaken by John Swan. The land immediately south of Tocal, along the west bank of the Paterson River, was farmed by Swan, who by early 1823 had already cleared at least 53 acres of vegetation (Dangar, 1823). As the boundaries of these early farms had not being officially surveyed and demarcated, definitive property boundaries were uncommon. This was reflected in the frequent disputes that occurred between neighbours about the ownership of the lands where two or more properties met (Hunter, 1997, p. 5). Where there was no adjacent land-owner, it is conceivable that a farmer's property could easily expand beyond its original grant.

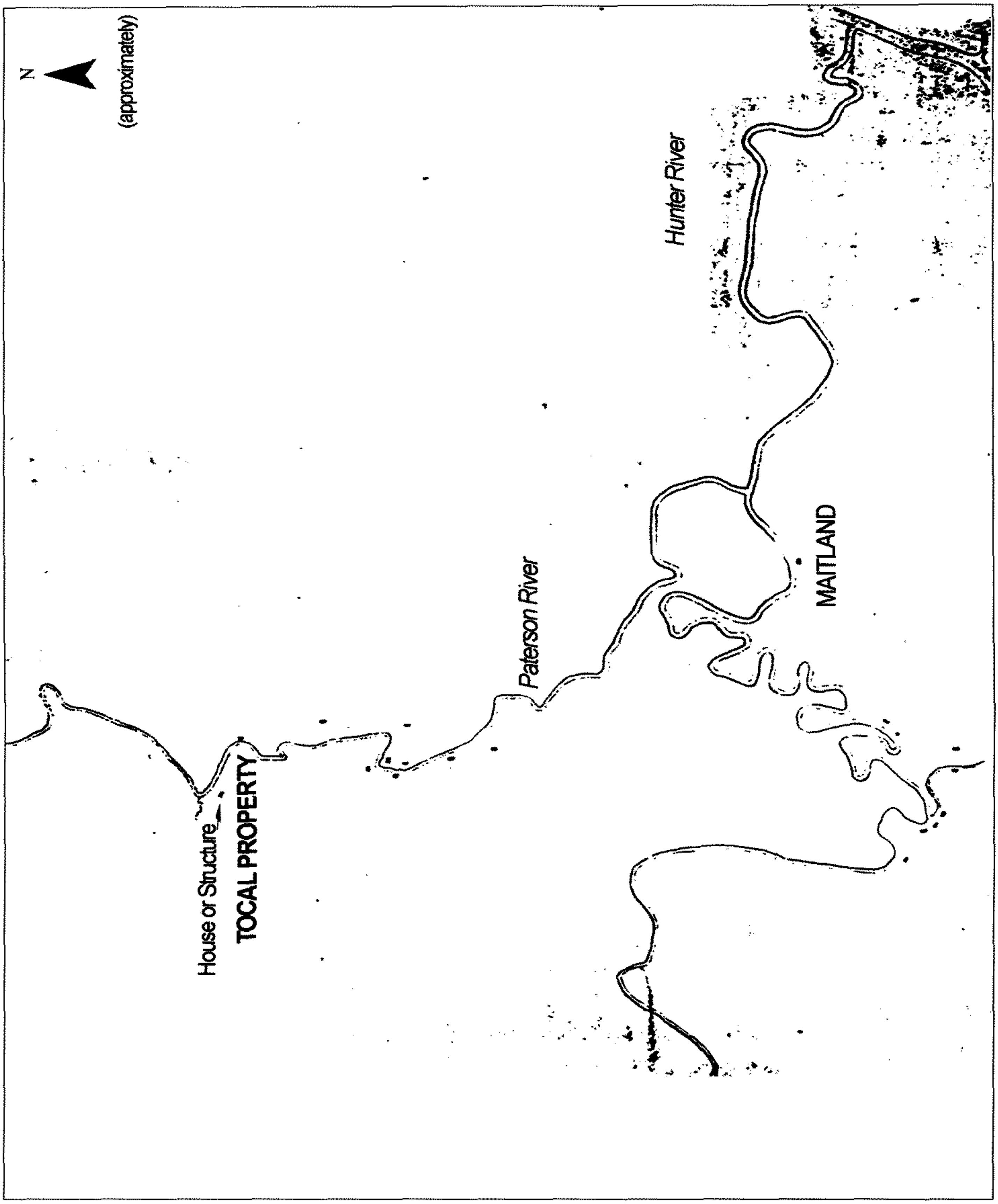


Figure 4.4 A map of the lower Hunter valley that is considered to predate Dangar's survey. The map has been redrawn from that reproduced in Hunter (1997), and shows that a structure existed in the Tocal catchment at this time.

John Swan's agricultural activities may therefore have extended beyond the limits of his grant into the adjacent northern land of Tocal, which at the time was without title. The unnamed dwelling in the pre-1823 map, which Dangar listed as an 'abandoned settlement' in his survey, may in fact represent the northern extent of the farming activities of John Swan.

Whether George Pell or John Swan is the owner of the unnamed dwelling in the pre-1823 map of Tocal, these several accounts all seem to suggest that land-clearing and farming was taking place in the southern portion of the Tocal site prior to the first official land grant in 1822.

4.4 The first land grant and the beginning of agriculture, 1822–1834 AD

4.4.1 The first European settlement at Tocal

With the opening of the land route from the Hawkesbury to the Hunter in 1819 by John Howe (Campbell, 1928), making the valley even more permeable to escaped convicts, and with increasing pressure to settle the rich lands along the Hunter, Governor Macquarie moved to transfer the Newcastle gaol further north to Port Macquarie. Permission to do this was received from London in 1820, allowing the lands adjacent to the Hunter River to be occupied by free settlers (Macquarie to Bathurst, 1819a; 1819b; Bathurst to Macquarie, 1820).

One such settler was James Phillip Webber, who arrived in Australia in early January 1822. On the 16 January 1822, Webber was notified by the Colonial Secretary that his request for a land grant of 1500 acres, submitted to the Colonial Office in London, had been received. Webber travelled to the Paterson River and by January 1822 had selected his grant on the west bank of the river. By mid-March, Webber and four convicts had moved onto his 1500 acre grant (Webber, 1822), which included the greater part of the Tocal catchment. A convict land-clearing gang was employed to deforest at least 50 acres (Walsh, 1999). Webber's original request for 1500 acres wasn't granted until September 1822, though for an enlarged area of 2000 acres. The new property was named *Markham* (Archer, 1988). In 1825, Webber applied for and was granted a further 1280 acres which adjoined the original grant to the west and north (White, 1986). Figure 4.5 shows the location of the original grant on the Paterson River.

4.4.2 The introduction of European agriculture at Tocal

Webber was an active agriculturist, and the Tocal property was extensively cultivated from 1822 onwards. During his ownership wheat, corn, tobacco, melons, bananas and stone fruit were all grown at various times. Viticulture was practised and Webber kept sheep and cattle, and maintained a dairy (Mitchell, 1984). The November 1828 Census of New South Wales recorded the stock, convicts kept and the number of acres cleared and cultivated for the site (Table 4.2).

Table 4.2 Property data for Tocal in 1828 (from the 1828 Census of New South Wales).

	Count
Convicts	38
Horses	3
Cattle	222
Sheep	1674
Cleared Acres	160
Cultivated Acres	120
Total Acres	3280

4.4.3 Early construction and dwellings

The first structures to be erected at Tocal were most likely impermanent wattle and daub huts, none of which remains today. Walsh (1999), using Cunningham's (1827) observations of convict housing in the area, has estimated that four or five huts may have been initially built. They were probably situated on the hill overlooking Tocal Homestead Lagoon and Webber's Creek, as this is where subsequent constructions were concentrated. The earliest remaining structure at Tocal today is a stone and wattle and daub dwelling known as Webber's Cottage (Figure 4.6), which is thought to have been built as early as 1822 (Walsh, 1999, p. 30). Dangar (1823) recorded visiting James Webber at 'his house' in 1822 as part of his survey of the region, suggesting that the cottage existed by this time. Structures that were erected between 1827 and 1830 are listed in Table 4.2, and are a good indication of the nature and scale of agricultural pursuits during the period. Not included in this list is a substantial stone barn that Webber built in the later part of 1830 (Table 4.3 and Figure 4.7), and three recently discovered subterranean silos (Figure 4.8).

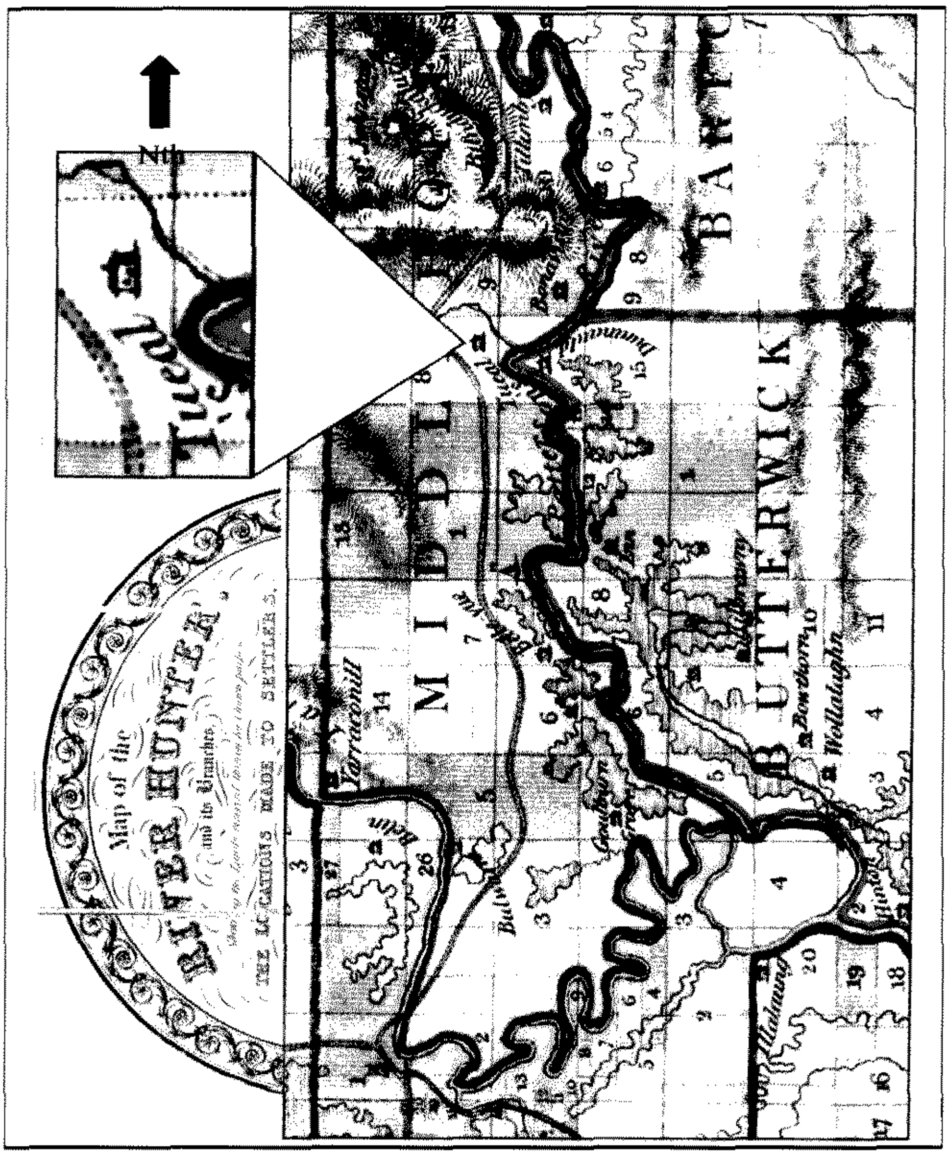


Figure 4.5 The location of Webber's original grant on the Paterson River near its confluence with Webber's Creek, from Dangar (1828).

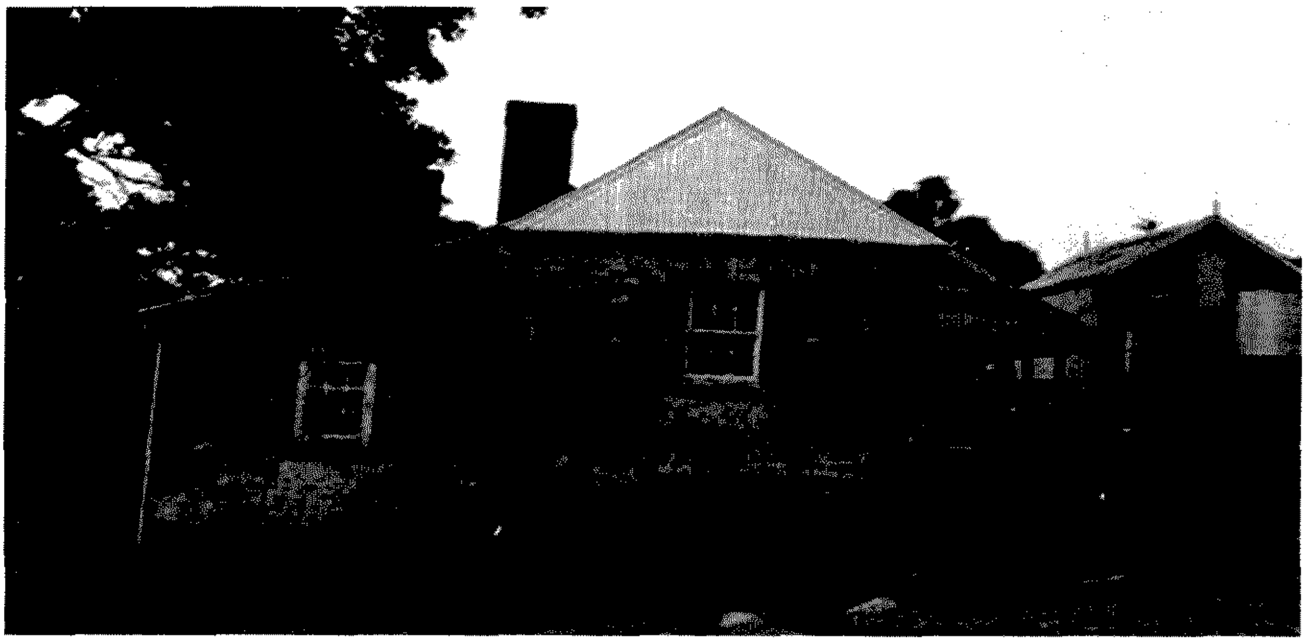


Figure 4.6 Webber's Cottage at Tocal, Paterson, eastern Australia, 1999.

Table 4.3 Dwellings and structures listed for valuation at Tocal, adapted from Walsh (1999).

October 1827	November 1829	May 1830
House	House	House
Barn	Barn	Barn
Maize barn	Dairy, store and granary	Dairy, store and granary
Kitchen	Kitchen	Tobacco sheds
House for servants	Tobacco sheds	Pressing houses
Two small servants houses	Two pressing houses	Maize barns ad smaller buildings
	Maize barns and smaller buildings	

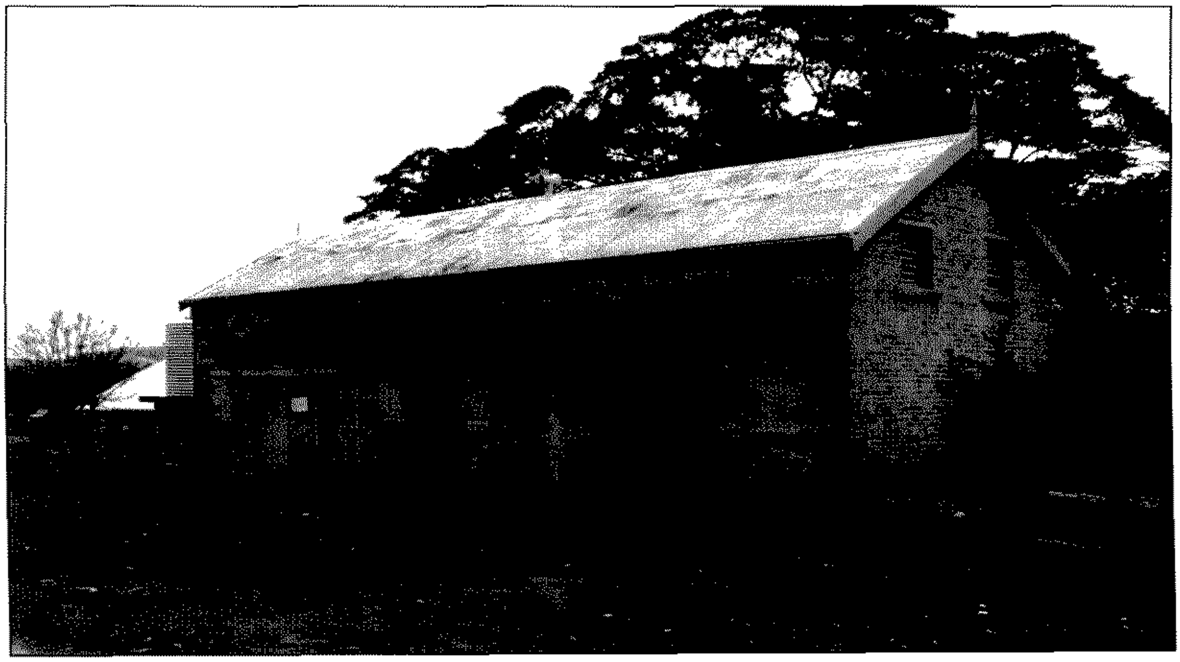


Figure 4.7 Webber's stone barn built in 1830, Tocal, Paterson, eastern Australia, 1999.

Figure 4.8 C.B. Alexander College students in one of the three large subterranean silos recently discovered in the Tocal catchment. The silos are thought to date from the 1830s. (Source: Tocal website (http://www.tocal.com))

4.4.4 The location of early agriculture in the Tocal catchment

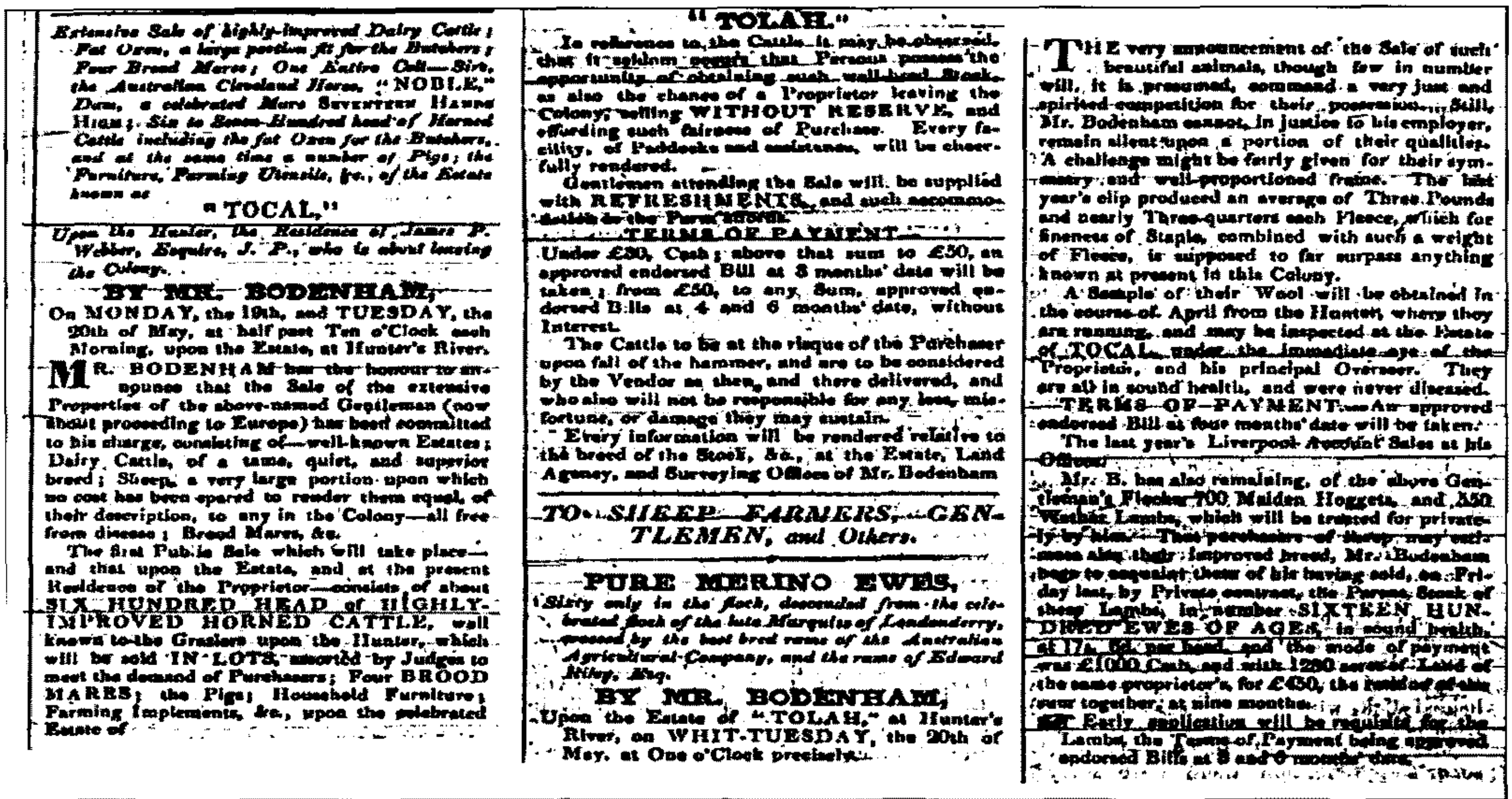
There is little information on the location of agricultural activities in the catchment during this period, though some insight comes from evidence preserved in the landscape today. The wine industry at Tocal involved the construction of raised mounds on the lower slopes in the southwest of the catchment. Records reported in Sullivan (1997) list Webber's vineyard as occupying three acres of land at Tocal. The historian Mitchell (1984) referred to Webber's sheep being kept in a paddock that ran parallel to the Paterson River. However, no primary source of information is listed for this record. The only other information on the location of agricultural pursuits at Tocal during the first years of European settlement comes from the location of ruins of early colonial structures. Remains of a bridge and a sheep wash in Webber's Creek, immediately north of the homestead suggest that many of the early farming activities were focused on the banks of the Creek. White (1986) suggests that Webber's crops were grown in the alluvial soils adjacent to Webber's Creek, and that Webber's vineyard was located on the slopes adjacent to the homestead.

By 1829, Webber had at least 2200 sheep at Tocal (Table 4.4), which, by his own admission, his '... land was unable to maintain'. He therefore applied for a further 2000–2500 acres (detailed in a letter written by Webber [1830] and reproduced in Walsh [1999, Appendix 9]). Webber was granted an additional 2560 acres in 1831 (Walsh, 1999, Appendix 11), by which time his Merino sheep population had reached 3000. Webber's estate had also expanded with the leasing of 320 acres to the south of Tocal.

Table 4.4 Stock at Tocal, 1829, compiled from Walsh (1999) and sources therein.

Stock type	Count
Rams	26
Horses	3
Cattle	350
Sheep	2200

Further information on stock numbers and the intensity of agricultural activity during Webber's ownership comes from advertisements that appeared in the *Sydney Herald*, 25 March (for an auction of stock held on 20 May) and 20 October 1834 (for the sale of the property) (see Figures 4.9 and 4.10). The second of these provides valuable information about the site's soil, agricultural activity and buildings. The advertisement also refers to the sale of 1600 'EWES OF AGES' the previous Friday, which are included in Table 4.5, which presents the total stock offered for sale in 1834.



Extensive Sale of highly-improved Dairy Cattle; Fat Oxen, a large portion fit for the Butchers; Four Brood Mares; One Entire Colt—Sire, the Australian Cleveland Horse, "NOBLE," Dam, a celebrated Mare Seventeen Hands High; Six to Seven-Hundred head of Horned Cattle including the fat Oxen for the Butchers, and at the same time a number of Pigs; the Furniture, Farming Utensils, &c., of the Estate known as

"TOCAL,"

Upon the Hunter, the Residence of James P. Webber, Esquire, J. P., who is about leaving the Colony.

BY MR. BODENHAM,

On MONDAY, the 19th, and TUESDAY, the 20th of May, at half past Ten o'Clock each Morning, upon the Estate, at Hunter's River.

MR. BODENHAM has the honour to announce that the Sale of the extensive Properties of the above-named Gentleman (now about proceeding to Europe) has been committed to his charge, consisting of—well-known Estates; Dairy Cattle, of a tame, quiet, and superior breed; Sheep, a very large portion upon which no cost has been spared to render them equal, of their description, to any in the Colony—all free from disease; Brood Mares, &c.

The first Public Sale which will take place—and that upon the Estate, and at the present Residence of the Proprietor—consists of about SIX HUNDRED HEAD of HIGHLY-IMPROVED HORNED CATTLE, well known to the Graziers upon the Hunter, which will be sold IN LOTS, assorted by Judges to meet the demand of Purchasers; Four BROOD MARES; the Pigs; Household Furniture; Farming Implements, &c., upon the celebrated Estate of

"TOLAH."

In reference to the Cattle it may be observed, that it seldom occurs that Persons possess the opportunity of obtaining such well-bred Stock, as also the chance of a Proprietor leaving the Colony, selling WITHOUT RESERVE, and affording such fairness of Purchase. Every facility, of Paddocks and assistance, will be cheerfully rendered.

Gentlemen attending the Sale will be supplied with REFRESHMENTS, and such accommodation as the Farm affords.

TERMS OF PAYMENT.

Under £30, Cash; above that sum to £50, an approved endorsed Bill at 3 months' date will be taken; from £50, to any Sum, approved endorsed Bills at 4 and 6 months' date, without Interest.

The Cattle to be at the risque of the Purchaser upon fall of the hammer, and are to be considered by the Vendor as then, and there delivered, and who also will not be responsible for any loss, misfortune, or damage they may sustain.

Every information will be rendered relative to the breed of the Stock, &c., at the Estate, Land Agency, and Surveying Offices of Mr. Bodenham

TO SHEEP FARMERS, GENTLEMEN, and Others.

PURE MERINO EWES,

Sixty only in the flock, descended from the celebrated flock of the late Marquis of Londonderry, crossed by the best bred rams of the Australian Agricultural Company, and the rams of Edward Riley, Esq.

BY MR. BODENHAM,

Upon the Estate of "TOLAH," at Hunter's River, on WHIT-TUESDAY, the 20th of May, at One o'Clock precisely.

THE very announcement of the Sale of such beautiful animals, though few in number will, it is presumed, command a very just and spirited competition for their possession. Still, Mr. Bodenham cannot, in justice to his employer, remain silent upon a portion of their qualities. A challenge might be fairly given for their symmetry and well-proportioned frame. The last year's clip produced an average of Three Pounds and nearly Three-quarters each Fleece, which for fineness of Staple, combined with such a weight of Fleece, is supposed to far surpass anything known at present in this Colony.

A Sample of their Wool will be obtained in the course of April from the Hunter, where they are running, and may be inspected at the Estate of TOCAL, under the immediate eye of the Proprietor, and his principal Overseer. They are all in sound health, and were never diseased.

TERMS OF PAYMENT.—An approved endorsed Bill at four months' date will be taken.

The last year's Liverpool Account Sales at his Offices.

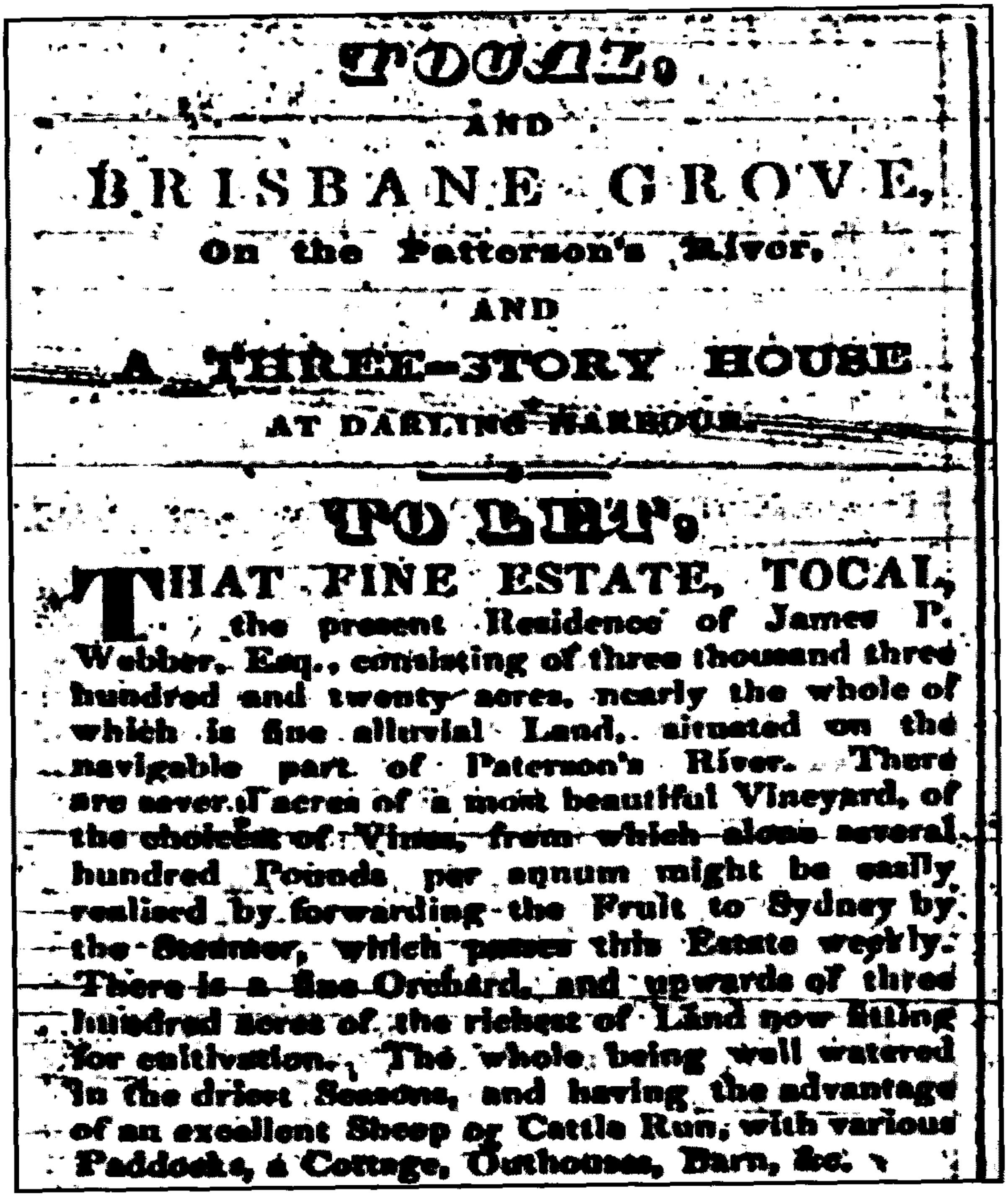
Mr. B. has also remaining, of the above Gentleman's Flocks 700 Maiden Hoggets, and 350 Wether Lambs, which will be treated for privately by him. That purchasers of sheep may estimate their improved breed, Mr. Bodenham begs to acquaint them of his having sold, on Friday last, by Private contract, the Parent Stock of these Lambs, in number SIXTEEN HUNDRED EWES OF AGES, in sound health, at 17s. 6d. per head, and the mode of payment was £1000 Cash, and with 1280 acres of Land of the same proprietor's, for £450, the residue of the [illegible] together, at nine months.

Early application will be requisite for the Lambs, the Terms of Payment being approved endorsed Bills at 3 and 6 months' date.

Figure 4.9 Advertisement from the *Sydney Herald* newspaper, 24 March 1834 for the sale of stock from Tocal.

4.4.5 The immediate post-settlement environment at Tocal and surrounds

A valuable source of information on the environment in the Tocal area in the 1820s is from the records of Cunningham (1827). His general impression of the soils along the Paterson and Hunter Rivers was that they were of 'amazing fertility'. These sentiments were repeated frequently in descriptions of the region's soils, although the soils that he observed on the hills away from the river were, 'inferior soil[s]'. Cunningham also commented on the physical characteristics of soils along the Paterson, describing one low relief, alluvial soil as a, 'poor washed clay or light sandy soil' (Cunningham, 1827). These descriptions are probably to be the first written record of the nature of soils in the region. In the same journal, there are two records by Cunningham of the presence of gullies formed along the river banks. The first describes gullies along the Hunter River between Newcastle and present day Maitland, some of which were difficult enough that 'partial unloading of the drays' would be required to traverse them (Cunningham, 1827, p. 145). When travelling towards 'Iron-bark Hill', Cunningham noted the presence of 'difficult gullies' that had formed along a cart track. Although it is difficult to assess whether these gullies predated European settlement of the area or not, Cunningham's description suggests an association between the gullying he observed and the cart-tracks.



TOCAL,
AND
BRISBANE GROVE,
On the Patterson's River,
AND
A THREE-STORY HOUSE
AT DARLING HARBOUR.

TO LET.

THAT FINE ESTATE, TOCAL, the present Residence of James P. Webber, Esq., consisting of three thousand three hundred and twenty acres, nearly the whole of which is fine alluvial Land, situated on the navigable part of Paterson's River. There are several acres of a most beautiful Vineyard, of the choicest of Vines, from which alone several hundred Pounds per annum might be easily realised by forwarding the Fruit to Sydney by the Steamer, which passes this Estate weekly. There is a fine Orchard, and upwards of three hundred acres of the richest of Land now fitting for cultivation. The whole being well watered in the driest Seasons, and having the advantage of an excellent Sheep or Cattle Run, with various Paddocks, a Cottage, Outhouses, Barn, &c.

Figure 4.10 Advertisement from the *Sydney Herald* newspaper, 20 August 1834 for the lease of the Tocal property.

Table 4.5 Stock offered for sale at Tocal, 1834, from the *Sydney Herald*, 25 March and 20 October 1834.

Stock type	Numbers
Maiden hoggets	700
Wether lambs	550
Ewes of ages	1600
Horned cattle	600
Horses	4

One visual record of the environment in the Paterson region from this period comes the paintings of Conrad Martens (Figure 4.11). His watercolour 'View of Trevallyn' shows the extent of vegetation removal from the riparian zones of the Paterson River in 1837.

Figure 4.11 Conrad Martens' 'View of Trevallyn' (1837). By this date extensive clearing of the Paterson floodplains is evident (Source: State Library of New South Wales).

4.5 Tocal and the Wilson Family, 1834–1844 AD

4.5.1 The Wilson family, fire and Tocal Homestead

Tocal was sold to Caleb and Felix Wilson in 1834 (Mitchell, 1984). In 1835 fire broke out in the vicinity of the lagoon resulting in the destruction of the various wooden outbuildings at the site (Anon., 1835). Several years later, Felix Wilson began work on the present homestead, which was completed in either 1840 or 1841 (Walsh, 1999). The National Trust registered building is situated on the western slope of the lagoon and is a significant tourist attraction in the lower Hunter valley. It has been suggested that the building was constructed of bricks formed from the mud of nearby Tocal Homestead Lagoon (Mitchell, 1984, p. 165), though most authors agree that the building is in fact constructed of sandstock brick and sandstone blocks, with the stone probably having been quarried from a site in the west of the catchment (White, 1986; Hunter, 1997; Walsh, 1999).

4.5.2 Agricultural expansion and its impact on the environment

From 1834 to 1844 agricultural pursuits continued with particular attention being devoted to viticulture, with Wilson producing award winning wines (White, 1986). During this period, Wilson further increased the size of the Tocal estate by purchasing three lots, each of 80 acres, adjacent to the site's southern boundary. Evidence of the intensity of agricultural activity at this time comes from the observations of Davidson (1846), whose accounts suggest that some farmers on the Paterson River may have been over-cropping their soils: 'Some farmers sow wheat on land from which they have just reaped a crop of Indian corn ...'. The effects of increasing soil erosion during the first decades of European agriculture are also recorded in accounts of the rapid sedimentation of the region's rivers. Observations of the extent of agricultural development along the Paterson in the 25 December 1841 issue of *The Hunter River Journal* recorded that along the 25 mile navigable length of the channel:

> ... and far beyond, the country on its banks is altogether reclaimed, and in a high state of cultivation.

A year earlier (1840) the extensive deforestation and intensive land-use resulted in the Paterson River silting up to the extent that, on occasion, the channel became impassable to the steam boats that had been the sole form of transport for produce within the area (Hunt, 1973, p. 10). Problems of soil infertility in the Paterson region were also becoming apparent. A column dedicated to 'rural intelligence' published in *The Hunter River Gazette* in late 1841 reported that 'In many localities the soil shows evident symptoms of exhaustion ...'.

4.6 The Reynolds family at Tocal, 1844–1926 AD

4.6.1 The shift towards pastoralism at Tocal

In 1844 Wilson leased Tocal to Charles Reynolds (White, 1986) signalling the beginning of the most significant shift in land use practices on the property since *c.* 1822. Reynolds initially stocked Tocal with 300 head of cattle and, in the following decades, Tocal became well known for raising prize-winning cattle and thoroughbred horses. Nevertheless, records show that Reynolds still grew crops at Tocal in the 1840s (though probably not as cash crops), and was improving the land by oxen and horse drawn ploughs (White, 1986). During the 1840s he was also one of the first farmers in the region to utilise a reaping and threshing machine (White, 1986). In 1850, Reynolds sold a range of stock used primarily for dairy or cropping activities (Anon., 1850). Although this sale was indicative of Reynolds' increasing concentrating on pastoralism, the soils of Tocal were still utilised for some cultivation over the next several decades. In August 1867 fire destroyed Blackett's Barn, located near Tocal Homestead. The event is well detailed in the 10 August 1867 *Maitland Mercury* (Anon., 1867), and shows that the fire, while intense, did not spread far beyond this single building.

A journalist from the *Town and Country Journal* visited Tocal in early 1870, and commented on the success of Reynolds' cattle ('the finest in the colony'), sheep and horses (Anon., 1870a). The article also referred to the crops grown at Tocal, which included tobacco, maize, oats, lucerne, potatoes and other vegetables. A further article in this periodical from 28 November, 1870 detailed the increased difficulty that Reynolds experienced in growing crops at Tocal, specifically wheat, and that in the future '... he has almost determined not to put any of the Tocal land under the plough' (Anon., 1870b). With the area under crops being reduced, the shift to stud farming had become even more significant. Reynolds acquired more land throughout the Paterson area in *c.* 1870 to accommodate his greatly increasing number of stock, while further clearing of the remaining catchment vegetation took place; an account of ring-barking of trees in 1871 is provided by White (1986, p. 55).

By 1871, only 200 acres, or approximately 6% of the Tocal site, was used for the cultivation of maize, oats, barley, pumpkins and lucerne (White, 1986, p. 57), with the remainder employed for stud activities. Hunt (1973) noted that Chinese market gardens were maintained at Tocal, though the primary source of this information cannot be found. He believed that they were located on the western slopes of Tocal Homestead Lagoon with water supplied to them from the lake by manual labour. Stock numbers continued to increase throughout the late 19th century, requiring further expansion of the property, which reached 5900 acres by 1900. Figure 4.12 is thought to be the oldest photograph of the lake, showing the Tocal herd grazing on the immediate slopes. At this time, Tocal's thoroughbred and cattle stock reached their zenith in terms of their numbers, popularity, prestige amongst the farming community and success in agricultural shows (White, 1986).

4.6.2 Environmental degradation at the beginning of the 20th century

Accompanying the increasing agricultural activity at Tocal and in the region was soil degradation and erosion, leading to the silting-up of the Paterson River. In 1902, J.H. Maiden, the Government Botanist of New South Wales, made the following comments highlighting the role of agriculture in accelerating erosion, as part of a discussion on the cause of increased flooding along the Hunter River (Maiden, 1902, pp. 110–111):

> … the innumerable sheep tracks are accentuated and the ground everywhere is pulverised by the feet of the sheep wondering after the scanty herbage. When the rain falls much of this pulverised soil, carrying with it grass plants (latent) and seeds of grasses and various forage plants must be washed into the creeks and again into the Hunter, which becomes discoloured … It is only a matter of a brief historical period when … these rich river and creek flats will find their way into the Pacific Ocean.

4.6.3 The Reynolds family at Tocal

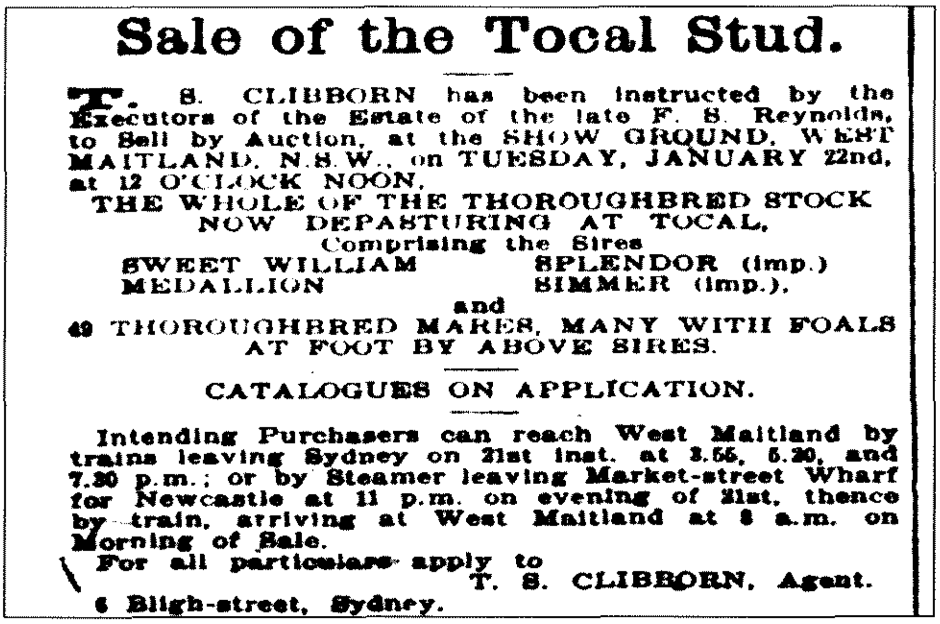
Charles Reynolds, the head of the Reynolds family, died in late 1871. Between 1844 and 1871, he changed the estate from a sheep and tobacco farm into an expansive stud-breeding empire (White, 1986). His wife Frances, with son Frank, continued to maintain the property until her death in 1900 (White, 1986). Much of the property's stock was sold on 22 January 1901, including all Tocal's thoroughbred horses (Figure 4.13).

Dry conditions through 1904–1905 were host to two large bushfires in the region of Tocal. The first of these, on 27 November 1904 broke out to the west of Tocal Homestead and north of Webber's Creek. Although it severely damaged many properties immediately north of Tocal ' … a strong muster from Tocal worked hard and after a few hours fighting succeeded in turning it off … ' its path southwards (Anon., 1904).

The second fire, on January 1 1905, was not so easily deflected. Accounts in the regional newspapers detail how, in the wake of the blaze, 'Mr Frank Reynolds estates Glendarra and Tocal were completely burnt out, only leaving the (Tocal) Homestead and stables.' (Anon. 1905a). Reynolds was considered by the Maitland Mercury's corresponded to be ' … a very heavy loser as most of the fencing was destroyed and not a blade of grass left for his large stock of cattle.' (Anon.,1905b).

Frank Reynolds purchased the Tocal property from the Wilson family in 1907, and began to restock Tocal with Hereford stock. Although Tocal was no longer one of the greatest Hereford studs in Australia, several articles were printed in the *Sydney Mail* (cited in White [1986]) through the 1910s praising the quality of Tocal's horse stock, while the Hereford stock continued to be well awarded at agricultural shows. Photographs from this period show the cattle and horses kept on the slopes of the lagoon, and in poorly vegetated paddocks adjacent to the homestead (Figures 4.14 and 4.15).

Figure 4.12 Tocal Homestead Lagoon *c.* 1900, with farm buildings and the Tocal Homestead in the background. This is one of the earliest images of agricultural activity in the Tocal Catchment, and demonstrates that cattle grazing was undertaken in this part of the Tocal property (Source: Tocal Library Archive).



Sale of the Tocal Stud.

T. S. CLIBBORN has been instructed by the Executors of the Estate of the late F. S. Reynolds, to Sell by Auction, at the SHOW GROUND, WEST MAITLAND, N.S.W., on TUESDAY, JANUARY 22nd, at 12 O'CLOCK NOON,

THE WHOLE OF THE THOROUGHBRED STOCK NOW DEPASTURING AT TOCAL,

Comprising the Sires

SWEET WILLIAM SPLENDOR (imp.)
MEDALLION SIMMER (imp.),

and

49 THOROUGHBRED MARES, MANY WITH FOALS AT FOOT BY ABOVE SIRES.

CATALOGUES ON APPLICATION.

Intending Purchasers can reach West Maitland by trains leaving Sydney on 21st inst. at 3.55, 5.20, and 7.30 p.m.; or by Steamer leaving Market-street Wharf for Newcastle at 11 p.m. on evening of 21st, thence by train, arriving at West Maitland at 8 a.m. on Morning of Sale.

For all particulars apply to

T. S. CLIBBORN, Agent.

6 Bligh-street, Sydney.

Figure 4.13 Advertisement for the sale of the Tocal Stud from the *Town and Country Journal* (Anon., 1901).

The branch line of the North Coast Railway from Maitland to Dungog opened in 1911. The railway was constructed through the middle of the Tocal property and involved extensive excavation and modification of the catchment's hydrology. Photographs taken at this time suggest that the amount of woodland cover at the site by 1911 to be roughly equal to that which remains today (Figure 4.16).

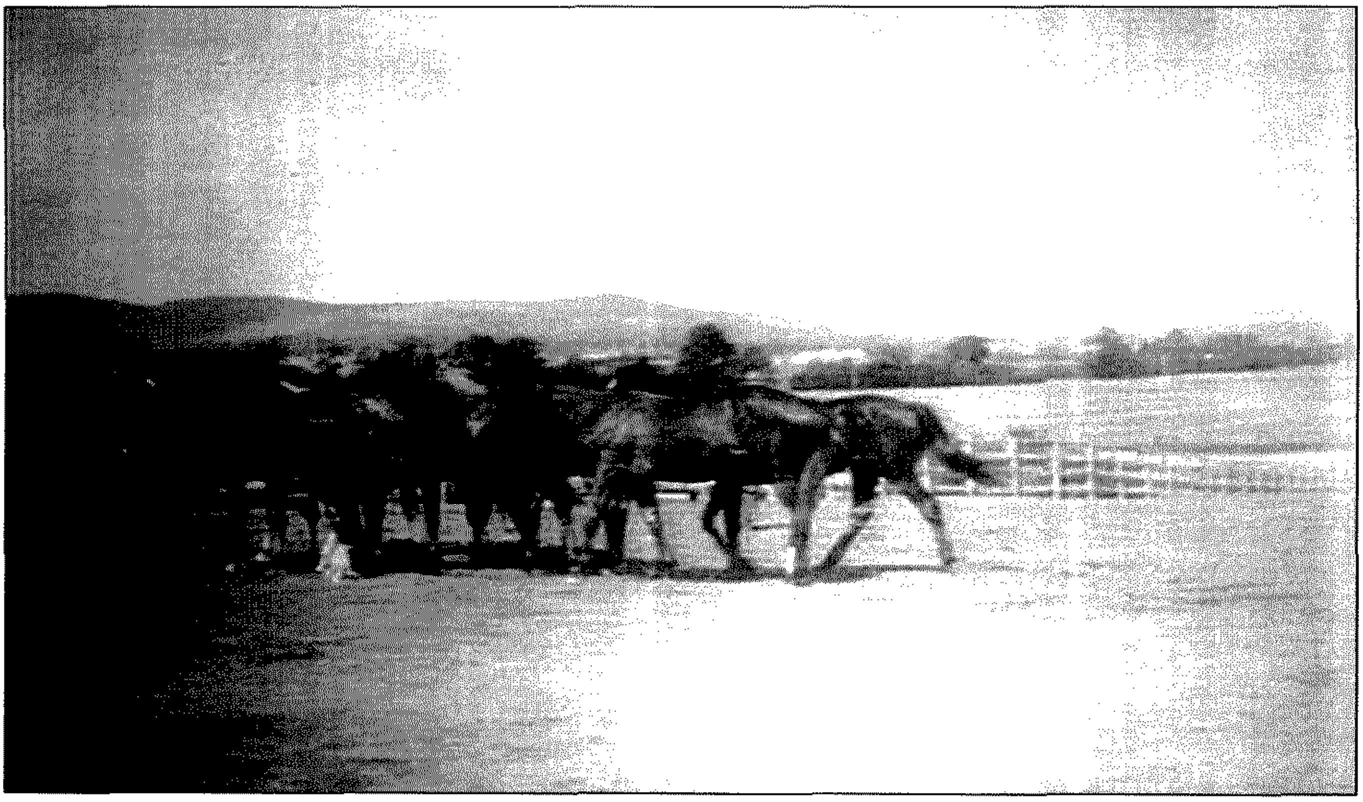


Figure 4.14 Horses in paddock on the western slope of Tocal Homestead Lagoon, 1910 or 1911 (Source: State Library of New South Wales).

Figure 4.15 Cattle at Tocal overlooking the lagoon, 1910 or 1911 (Source: State Library of New South Wales).

Figure 4.16 A racehorse at Tocal, early 1900s. The woodland visible in the background is of approximately the same extent as it is today. The photograph is taken looking south from the cluster of farm buildings (Source: Tocal Library Archives).

4.7 The early 20th century at Tocal

4.7.1 The Alexander family and Tocal

As late as 1926, stud farming was the predominant land use at Tocal. The Alexander family purchased the property from the Reynolds family in late 1926, and stocked the site with 101 head of cattle. By 1936, the Alexander family had shifted the main focus of the property to cattle grazing and fattening (White, 1986). Table 4.6 lists the sales of cattle from 1929 to 1937.

Table 4.6 Number of cattle sold from Tocal, 1929–1937. Data from White (1986).

Year	Number of cattle sold
1929	346
1930	860
1931	524
1932	589
1933	405
1934	680
1935	658
1936	824
1937	519

Through the first half of the 20th century agricultural activity generally declined. After 1927, thoroughbred horses were no longer kept at Tocal (White, 1986, p. 79). During the following decade, economic and environmental difficulties hampered agricultural productivity. Across much of the Australian continent, the early 20th century saw increased erosion due to rabbit infestations (Wasson and Clark, 1985; Gale and Haworth, 2005). Rabbit populations are thought to have been well controlled at Tocal, and the impact of rabbits on the site was probably of little importance (Mr A.C. Archer, principal, C.B. Alexander College, personal communication, 2005).

4.7.2 Drought and fire

Severe droughts were experienced across the region in the 1930s, while two uncontrolled fires damaged the Tocal property in 1939 (exact date unknown) and on 10 December 1944 (White, 1986, p. 90). The first of these was recorded as follows:

> ... a serious bushfire swept through Tocal, jumped the road [Tocal Road] and took hold of Glendarra. The homestead on Glendarra was saved but the loss of fences and pastures weighed heavily ...

The 1939 fire may have had little impact in the Tocal Homestead Lagoon catchment. The bushfire of five years later, however, was catastrophic. Much of the region, including most of the Tocal property and its buildings, was burnt. Mr Cameron Archer (principal, C.B. Alexander College, personal communication, 2005) has compiled diary entries and interviews with local residents which describe the fire of 1944. While these accounts suggest that most of the Tocal catchment was severely burnt, the Tocal Homestead and surrounding buildings were spared. It was this fire that was responsible for burning down the house 'Glendarra', located on the eastern edge of the catchment, taking with some of Tocal's historical and agricultural records. This suggests that the eastern part of the catchment may have been most affected.

By the 1940s, only one family member remained at Tocal, C.B. Alexander, and agricultural activities had essentially ceased. C.B. Alexander kept automobiles at Tocal from the 1930s onwards, these included 1929 and 1934 model Rolls Royces (White, 1986, p. 89). Alexander died in 1947 and although his two nieces occupied the Tocal homestead, the remainder of the property was unattended. This was due in part to legal difficulties associated with the implementation of C.B. Alexander's complex will for the future of the property. As a result, agricultural activity was effectively non-existent at Tocal for the following 15 years (White, 1986). This period finished in the early 1960s with the establishment of an agricultural college at Tocal.

4.8 The C.B. Alexander Agricultural College, 1965 to present

In 1964 work began on the campus buildings of the C.B. Alexander Presbyterian Agricultural College. A photograph from early 1964 shows the site to be overgrown by grasses, weeds and possibly shrubs (Figure 4.17) compared to today. The Presbyterian Church had taken control of the property the previous year, and some clearing of overgrowth and timber had already begun (White, 1986). E. A. Hunt, who was significantly involved in the conversion of the Tocal farm to an agricultural college, recorded many of the improvements that were made in the catchment following the previous 15 years of inactivity, such as these that took place in 1965 (Hunt, 1973):

> ... work had been concentrated on cleaning out old dams and building new ones ... draining low lying areas, felling timber, dealing with aftergrowth, and preparing for pasture improvements ...

Extensive vegetation clearing and modification of the catchment continued in 1966, and paddocks on the property were '... thoroughly cleared of lying timber, wire, rubbish and rocks ...', at least two dams were cleaned out and overgrown vegetation cleared (Hunt, 1973, pp. 49-50). The college opened in 1965, and agricultural activities became the main focus of the property once more. Information on the level of agricultural activity at Tocal during this period comes from an in-house annual magazine, *The Tocal College Magazine*. The 1966 edition lists the stock of the property at the time of the take over (Table 4.7) and reports the number of additional stock purchased that same year (Table 4.8). In this first year of the college, cattle numbers alone increased by 64%.

Figure 4.17 Construction workers at Tocal before commencing building of the college, February (?) 1964 (from the Tocal website (http://www.tocal.com).

Table 4.7 Stock at Tocal in 1965 at the time of its take over by the C.B. Alexander College, from Anon. (1966).

Cattle	Number	Horses	Number
Bullocks	139	Mare	1
Hereford bull	1	Filly, unbroken	1
Hereford cows	11	Gelding, unbroken	1
Hereford calf	1	Gelding, aged	1
Dairy cows	71		
Dairy calves	19		
Total cattle	**242**	**Total horses**	**4**

Table 4.8 Stock purchased in 1965 following the take over of Tocal by the C.B. Alexander College, from Anon. (1966).

Cattle	Number	Horses	Number
Dairy heifers	100	Horses	9
Hereford bulls	3		
Hereford breeders (unspecified)	52		
Total cattle	155	**Total Horses**	9

As part of its expansion of agricultural pursuits, the College acquired land to the west of the catchment in 1966 and a turf airstrip was constructed for the aerial distribution of agricultural chemicals, mainly superphosphate (Hunt, 1973). The first recorded use of the air strip was in late 1968 when 75 tonnes of superphosphate were applied to paddocks in the western portion of the property (Anon., 1968a).

Superphosphate fertilisers were first aerially distributed from 1950 onwards, starting on the New England Tablelands of New South Wales (Gale *et al.* [1995] and references therein). Aerial topdressing of agricultural land in the upper Hunter valley with superphosphate began in 1957 at a rate of 300 tons per annum, increasing to 20 000 tons per annum by *c.* 1969 (Boyle, 1969). Trucks delivered large amounts of superphosphate to a farm from where the fertiliser was loaded into the aircraft for distribution. Farms were then left with fertiliser 'dumps' in their paddocks (Figure 4.18) (Boyle, 1969).

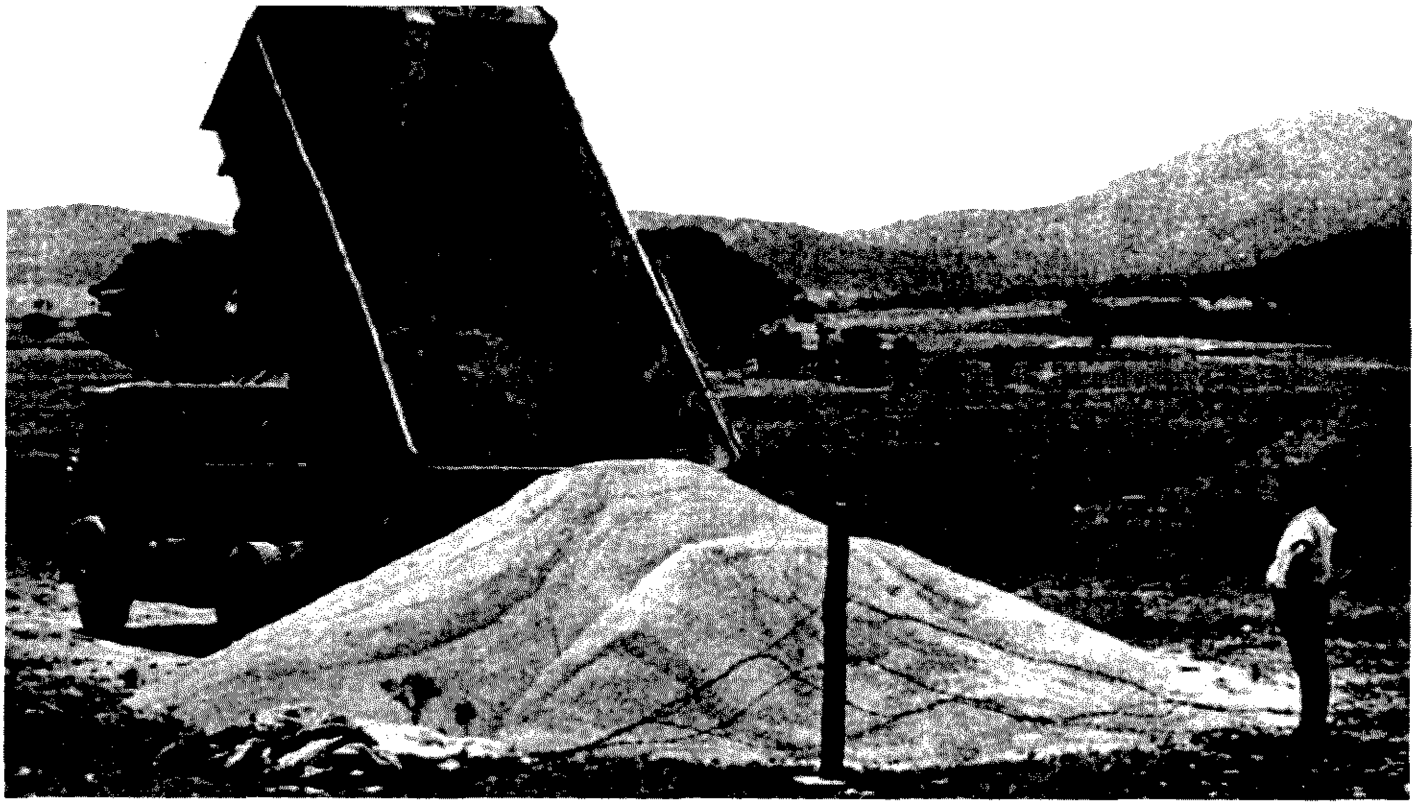


Figure 4.18 Superphosphate dump on farmland, upper Hunter valley, date unknown, from Boyle (1965).

The chemical compounds distributed included superphosphate, and gypsum- and sulphur-enriched superphosphates (known as SF30 and SF50) (Boyle, 1969). The timing and magnitude of the aerial distribution of agricultural chemicals at Tocal are likely to have been broadly similar to those experienced in the upper Hunter valley. The rapid increase in the use of aerially distributed superphosphate across New South Wales can be seen in the figures listed in Table 4.9.

Table 4.9 Aerial distribution of superphosphate in New South Wales. Adapted from Boyle (1969).

Year ending June	Area aerially topdressed and/or seeded (Acres)	Mass of superphosphate (Tons)
1957	917 895	45 329
1958	883 763	43 855
1959	1 048 410	48 791
1960	2 378 868	108 496
1961	3 969 052	192 893
1962	4 431 788	207 817
1963	5 521 768	253 213
1964	8 064 979	367 185
1965	10 074 894	463 584
1966	7 347 697	361 741
1967	8 058 000	335 153

Through 1967 and 1968 *The Tocal College Magazine* reported the further construction of campus buildings, significant increases in bovine stock numbers, and the aerial distribution of agricultural chemicals. Photographs of the construction of the college and facilities in the catchment show that extensive excavations and modification of the catchment took place during this period (Figures 4.19–4.21). Hunt (1972) noted that the construction of the college's playing fields and tennis courts alone required 'a great deal of cutting and filling' of the site.

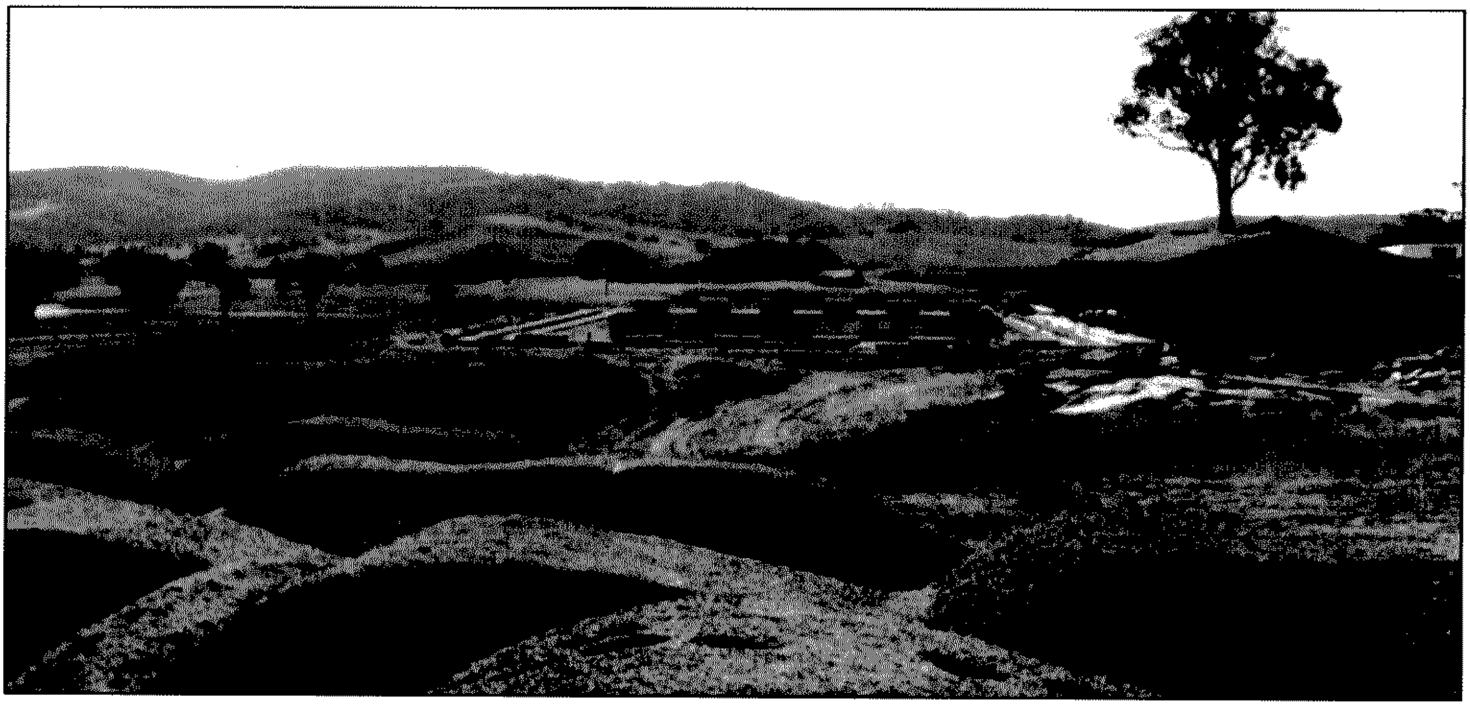


Figure 4.19 Photograph taken in 1966 during the construction of the Tocal college campus (Source: Tocal Library Archives).

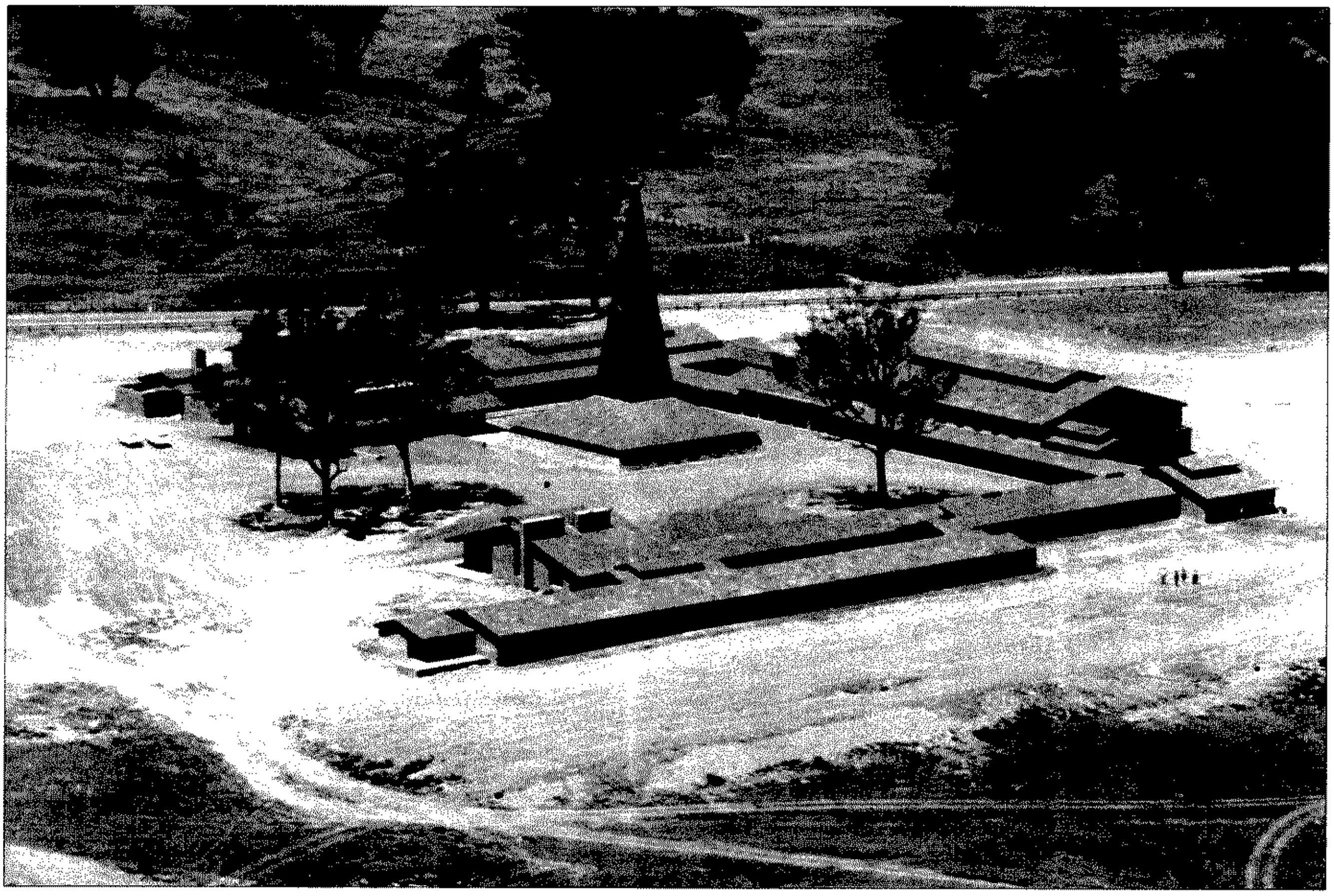


Figure 4.20 Oblique aerial photograph showing the newly completed college buildings *c.* 1966, from White (1986).

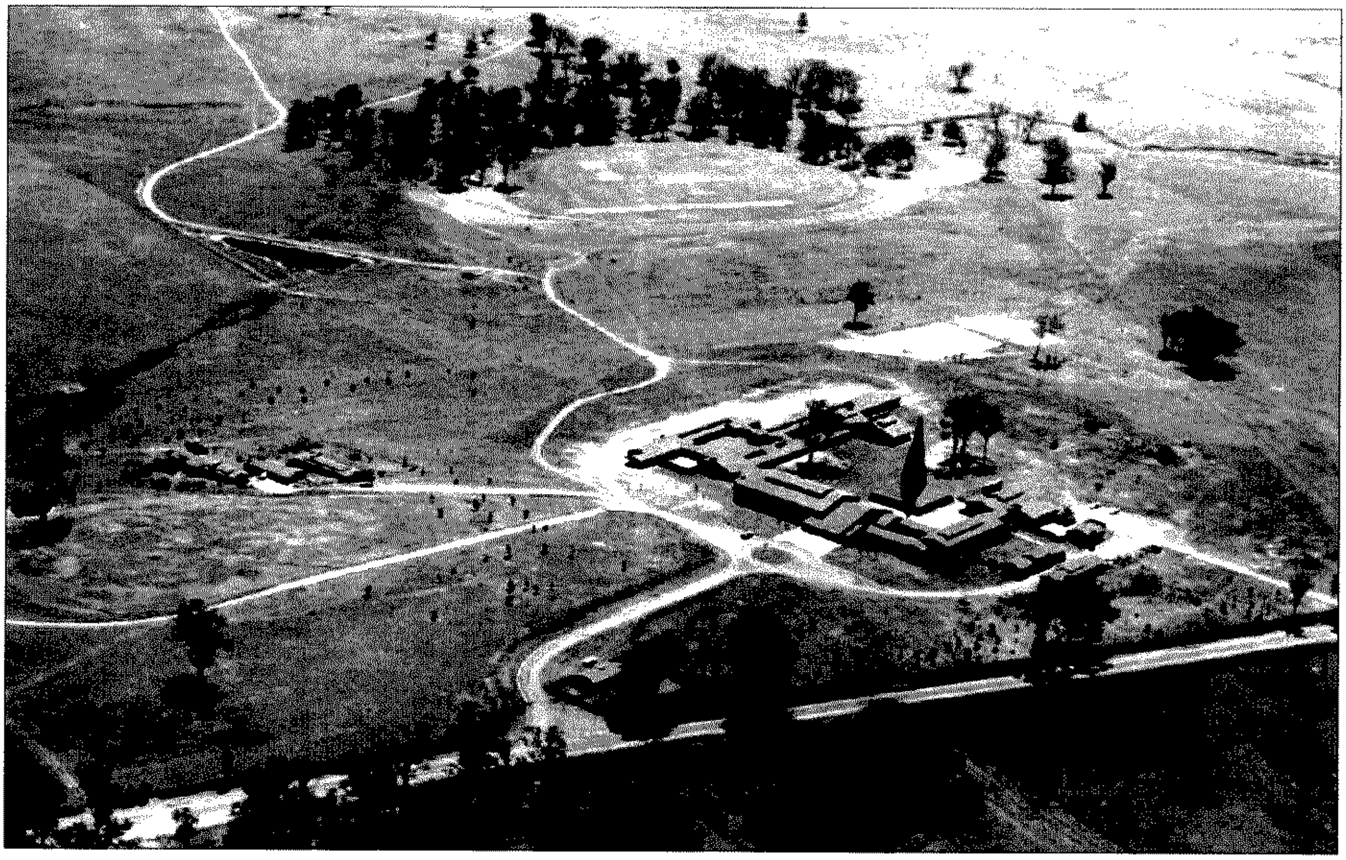


Figure 4.21 Oblique aerial photograph of the Tocal College, 1966 (Source: Tocal Library Archives).

In 1970 the Presbyterian Church transferred ownership of the College to the New South Wales Department of Agriculture. The College began acquiring more land during the late 1970s to accompany an expansion of its agricultural activities, and by 1981 the Tocal property encompassed 5265 acres (White, 1986). Although agricultural education had become the predominant activity of the Tocal property, the college maintained a poultry farm, and began a commercially orientated dairy in 1966 to supply products to the Hunter region.

Tourism and agricultural education are two other major activities at Tocal today. Up to 30 000 people visit Tocal for the annual Field Day, which was first held in 1984 (Figure 4.22). Tourists also arrive year-round, often on organised coach tours, to visit the heritage-listed Tocal Homestead, and the historic barns and stone cottages of the site.

In the last decade, more sustainable farming techniques have been adopted at Tocal and a *Tocal Code of Land Use Practice* (Anon., 1997) been published. The publication details past chemical pollution of the site, listing the location and nature of chemical storage sites in the southwest corner of the catchment, and the possibility of residual Dieldrin Super (an organochlorine pesticide since banned that had been used to control black beetle populations) in the soils. Issues of *The Tocal Catchment Magazine* (1968 and 1967) also noted the use of DTP for the large-scale clearing of vegetation at the site during the redevelopment of the property in the early 1960s.

In keeping with the recommendations made in the *Tocal Code of Land Use Practice*, the College, in collaboration with the Hunter Valley Catchment Management Trust, has undertaken large-scale remediation of the catchment, including the re-establishment of wetland areas, replanting of riparian vegetation, and participating in research into the effects of present day agricultural practices, such as the operation of the dairy (e.g., Geary and Moore, 1999). The adoption of more sustainable agricultural practices in the Hunter valley is thought to have already reigned in the amount of soil erosion (Lockie and Martin, 1993).

Figure 4.22 Oblique aerial photograph of the 1983 Tocal Field Day, which takes place in the southern portion of the catchment, from White (1986).

4.9 Soil erosion surveys in the Hunter valley

The Hunter valley of New South Wales was one of the first regions in Australia in which an organisation dedicated to monitoring soil erosion was established. The Hunter Valley Conservation Trust (now known as the Hunter Valley Catchment Management Trust) was established in 1950–1951, and was involved not only in surveying the region for data on soil erosion, but also in the design and funding of erosion remediation programs (Lockie and Martin, 1993). Four main surveys of soil erosion in the Hunter valley have been completed to date, 1941–1943 (Kaleski, 1945), 1967 (Stewart, 1968), 1970 (Higginson, 1973) and 1983–1985 (Emery, 1989). Kaleski's survey of 1942–1943 was accomplished by field mapping eight erosion classes (condensed later to six) and eight land classes at a scale of two miles to an inch (1:126 720) for regions in central and eastern New South Wales. He demonstrated that in the Hunter valley only 42% of land was unaffected by erosion (Table 4.10) (Kaleski, 1945, p. 15). Twenty five years later, Stewart (1968) undertook a new survey of soil erosion in part to test whether attempts at controlling erosion implemented after the Kaleski survey had been successful. To aid this comparison, Stewart adopted the same procedure used earlier by Kaleski, based on mapping and field observations. The results of the 1967 survey are presented in Table 4.11.

Higginson's 1970 survey used aerial photos to map seven erosion classes at scales ranging from 1:31 680 to 1:63 360 (Higginson, 1973). The survey defined 42 land systems and 10 land use classes (row crops, field crops etc.), and used cluster analysis to define erosion hazards for each land system based on its erosion classes (Table 4.12). The survey reported that 54.8% of the Hunter valley was unaffected by erosion, 25.2% was affected by sheet erosion and 20% was affected by gully erosion (of different severities). It concluded that most erosion was associated with grazing activities, usually on undulating to hilly land.

Table 4.10 The Kaleski (1945) survey of soil erosion in the Hunter valley, New South Wales.

Erosion class	Area in class (km^2)	Percentage of total area
Severe and extensive gully erosion	613.8	2.4
Moderate gully erosion	9041.7	36.2
Sheet erosion	4838.1	19.4
Moderate wind erosion	none	-
Severe wind erosion	none	-
No appreciable erosion	10463.6	42.0

The most recent survey of erosion in the Hunter valley, that by Emery (1989), was based on aerial photograph interpretation and estimates of erosion hazard made using the Universal Soil Loss Equation. Emery defined eight erosion classes, six based on area, and two classes for gully and streambank erosion based on the length of impacted channel (Table 4.13). The survey showed soil erosion to have increased dramatically, with only 19.6% of land unaffected, and sheet erosion affecting 79.1% of land. However, as Lockie and Martin (1993) noted, because erosion is predicted in this survey, rather than mapped from direct field observations, comparisons with earlier erosion survey data from the Hunter region may be inappropriate.

Table 4.11 The Stewart (1968) survey of soil erosion in the Hunter valley, New South Wales. The results are compared with those of the Kaleski (1945) survey.

Erosion class	Area (km^2)	Change since 1945	Percentage of total area	Percentage change since 1945 (%)
Severe and extensive gully erosion	525.8	-88.0	2.1	-0.3
Moderate gully erosion	9212.6	+170.9	36.9	+0.7
Sheet erosion	1955.5	2882.6	7.8	-11.6
Moderate wind erosion	none	-	0.0	0.0
Severe wind erosion	none	-	0.0	0.0
No appreciable erosion	13263.4	+7200.2	53.2	+11.2

Table 4.12 The Higginson (1973) survey of soil erosion in the Hunter valley, New South Wales.

Erosion class	Area (km^2)	Percentage of total area
Very severe gully erosion	44	0.2
Severe gully erosion	262	1.3
Moderate gully erosion	1619	7.8
Minor gully erosion	2222	10.7
Moderate to severe sheet erosion	197	0.9
Minor sheet erosion	5044	24.3
No appreciable erosion	11 362	54.8

Table 4.13 The Emery (1989) survey of soil erosion in the Hunter valley, New South Wales.

Erosion class	Area (ha)	Percentage of total area
Sheet erosion	1 655 737	79.1
Rill erosion	10 428	0.5
Wind erosion	1541	0.1
Mass movement	17 084	0.8
No appreciable erosion	409 591	19.6
Total erosion	1684 789	80.4
Erosion class	**Length (km)**	**Density (km/km^2)**
Gully erosion	16 109.5	0.769
Streambank erosion	1941.4	0.093

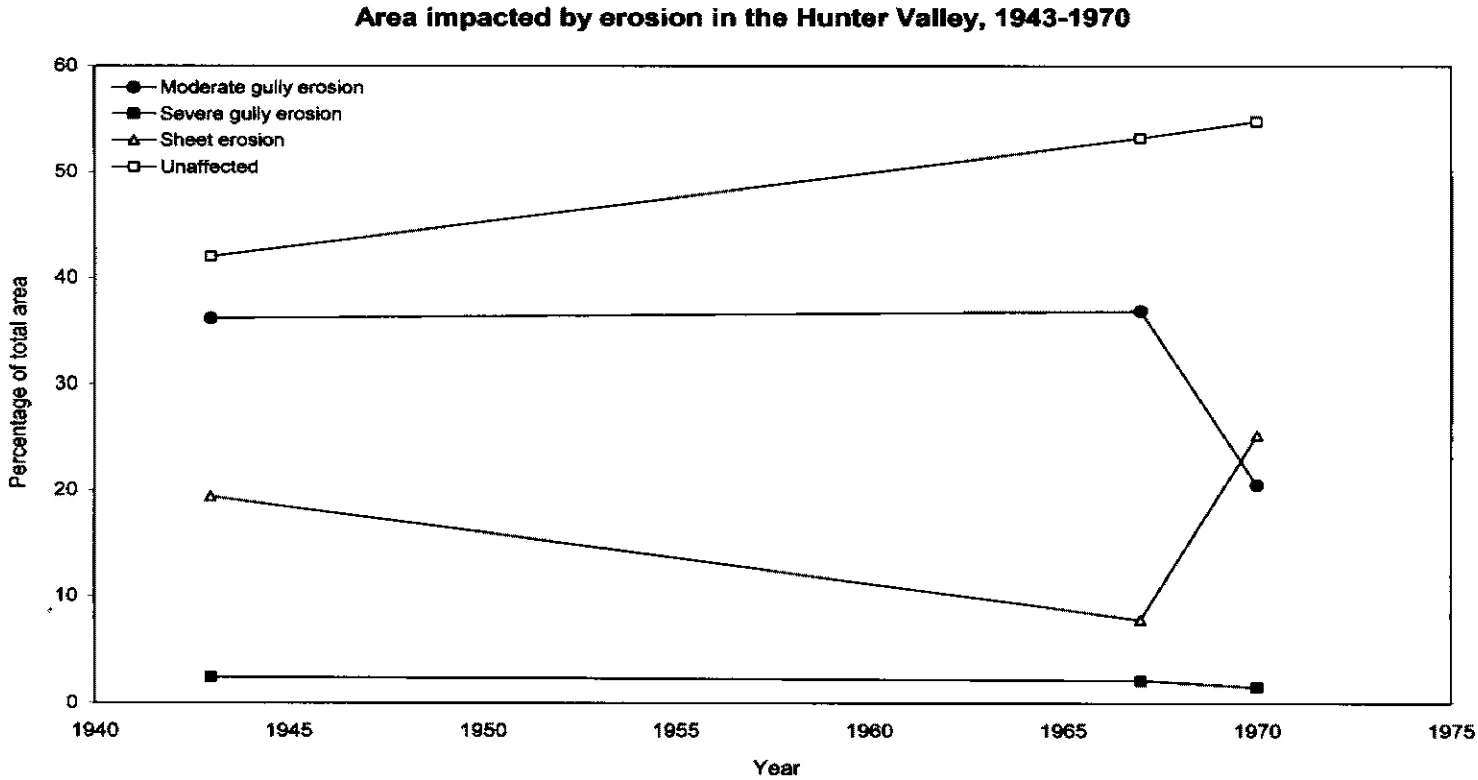


Figure 4.23 The area affected by erosion in the Hunter valley of New South Wales from 1943 to 1970.

4.10 Summary

The catchment of Tocal Homestead Lagoon has been utilised for a variety of agricultural-based activities since at least 1822, though some deforestation of the riparian areas of the site may have taken place up to a decade earlier. Although the identity of the catchment's first European settler is unknown, it is possible that European occupation of the site had taken place prior to the land being granted to James Webber in early 1822. Unfortunately, there is no information available from the historical or archaeological record on pre-European land-use by Aboriginal people, or on population densities or the length and continuity of their occupation of the site.

During the 19th century, the predominant land-use at Tocal shifted from cropping and sheep grazing to cattle stud activity, while during the last half century the site has been used for a mix of dairying, sheep grazing and agricultural education under the direction of the C.B. Alexander Agricultural College. Unfortunately, absolute stock densities for much of the historical period cannot be determined, due to the loss of many of Tocal's agricultural records in the bushfire of 1944 (Cameron Archer, principal, C.B. Alexander College, personal communication, 2005). However, information quarried from the historical record has allowed the reconstruction of the variations in patterns of land-use through this period. These are summarised in Figure 4.24. Such variations in land-use are likely to have had equally variable impacts on the catchment.

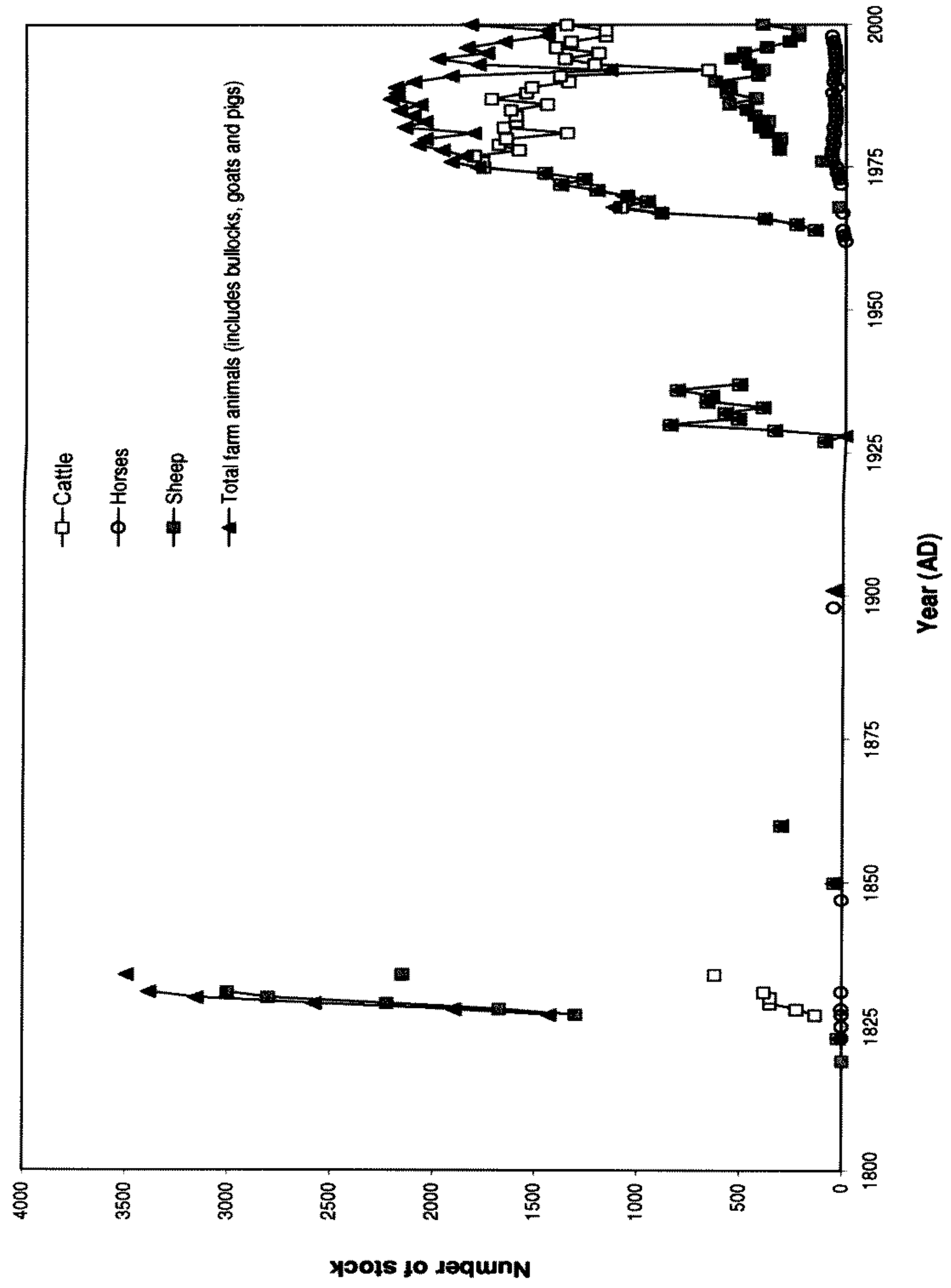


Figure 4.24 Changes in animal populations at Tocal, 1822 to 1998. Data taken from a wide range of sources cited throughout this chapter.

Chapter Five

Methods

5.1 Survey of the lake and catchment

5.1.1 Survey methods

Preliminary mapping of the Tocal Homestead Lagoon catchment was undertaken in 1998 using a combination of field survey, aerial photograph interpretation and published 1:25 000 topographic maps. Catchment boundaries, geology and building locations were surveyed using Trimble single-frequency GPS Total Station equipment. The perimeter of the lake had recently been surveyed by Scott and Crisp Pty Ltd Surveyors (1998) and these data were integrated into the project. Between 1999 and 2001 a more detailed survey of the catchment was undertaken with the aim of accurately defining the catchment boundary of Tocal Homestead Lagoon, capturing high-resolution contour data of the catchment and placing the data within the Australian Map Grid (AMG84) system for mapping.

5.1.2 Catchment hydrology and drainage density

Investigating changes in catchment drainage density (the total catchment area divided by the total length of channels in the catchment) over time is a well established means of assessing erosional processes over historical time spans (Strahler, 1958). Changes in catchment drainage density provide information on whether the drainage network is expanded or contracting and may provide information on a catchment's response to variations in land-use. Stream lengths were measured (in a horizontal plane) from the digitized aerial photographs in the Tocal GIS for the following years: 1958, 1967, 1984 and 1996. The drainage basin of the lake was measured (again, in a horizontal plane) using the catchment survey data.

5.2 Lake sediment sampling

A 50 m grid sampling strategy was devised for the lake based on transects aligned north–south, parallel with the main axis of the Tocal Homestead Lagoon basin (Figure 5.1). Twenty-two cores were extracted from the lake by percussion coring from a motorised pontoon (Figure 5.2). All 22 of the cores were taken using 50 mm diameter polyvinyl chloride (PVC) tubing (Figure 5.3). This allowed the subsequent measurement of downcore variations in volume-specific magnetic susceptibility without extruding the cores from their sleeves. The average horizontal sampling density across the entire lake basin was approximately 1 in 0.3 ha, while the sampling density within the 50 m sampling grid was nearly 1 in 0.06 ha. These sampling densities lie between the middle (1 in 0.3 ha) and the higher end (1 in 0.06 ha) of the range of sampling densities common to whole-basin studies of sedimentary flux world-wide (Dearing and Foster, 1986).

5.3 Core logging and sub-sampling

5.3.1 Volume-specific magnetic susceptibility

The volume-specific magnetic susceptibility (VSMS) of each core was measured using a Bartington MS2C core scanning susceptibility sensor interfaced with a Bartington magnetic susceptibility meter Model MS2 (Thompson and Oldfield, 1986, pp. 105–106; Gale and Hoare, 1991, pp. 222–223). By aligning 'peaks' and 'troughs' common to each downcore sequence (or curve matching), trends in magnetic signals thought to be contemporaneous at each coring location were identified across the entire basin. This allowed the construction of a magnetostratigraphy spanning the entire lake basin, following the procedure outlined by Dearing (1986).

5.3.2 Additional cores from location TCA9

The core location that yielded the longest, highest resolution and most complete volume-specific susceptibility record, TCA9, was chosen for additional sampling. A master core (TCA9b) was taken from this location on 17 June 1998 using 101 mm diameter steel tubing, enabling the recovery of 3.523 m of sediment. This core provided the greatly increased volume of material required for dating, geochemistry, pollen analysis and more detailed magnetic analyses.

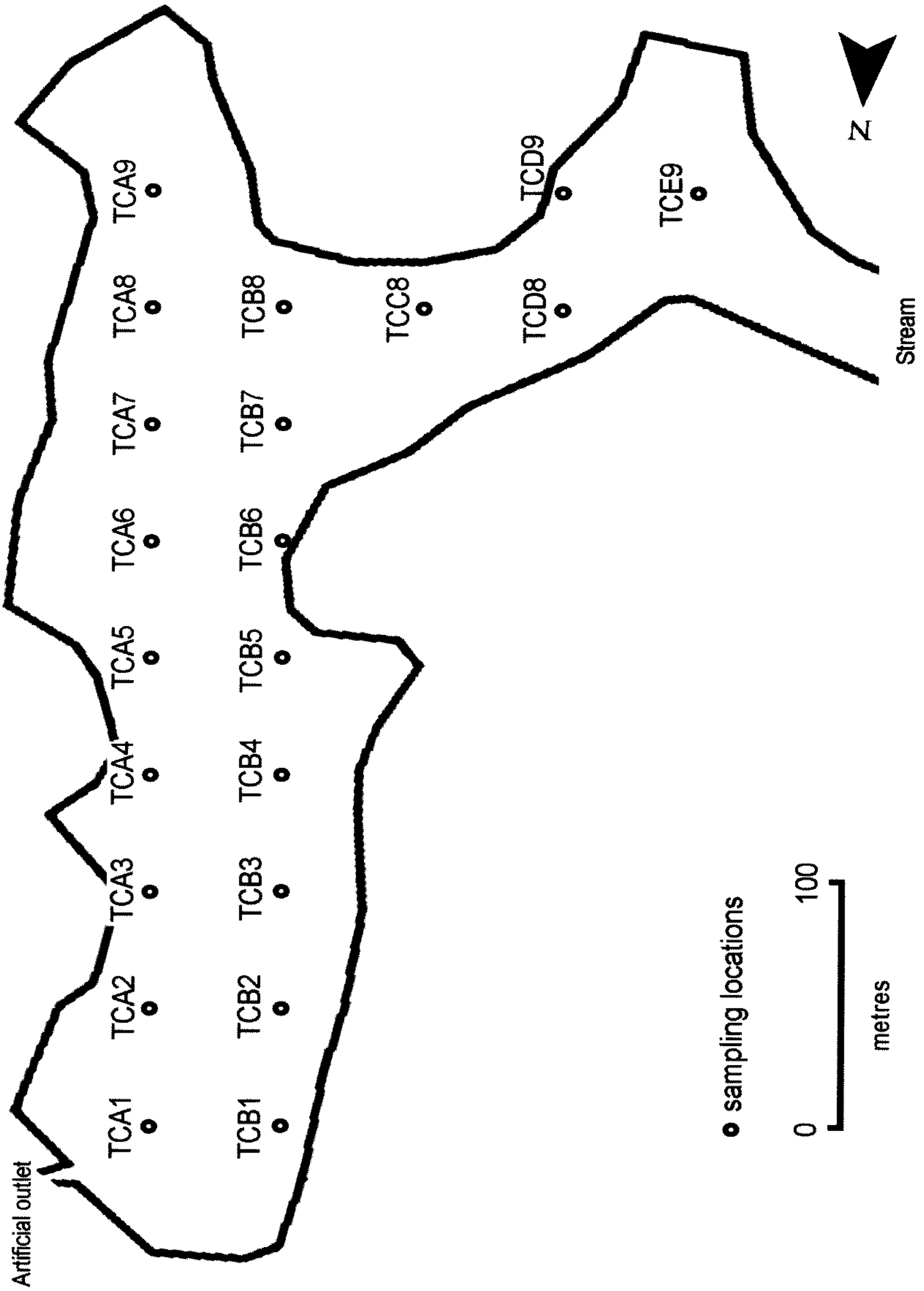


Figure 5.1 Locations within Tocal Homestead Lagoon from which 50 mm diameter sediment cores have been extracted.

In 2004, two further 50 mm PVC cores of 6 m length were extracted from location TCA9 to provide material for palaeomagnetic analysis (section 5.4.8). The direction of magnetic north was recorded on each core barrel prior to their extraction from the lake.

5.3.3 Sub-sampling and measurement of physical properties, core TCA9b

The master core was opened by making vertical cuts along opposite sides of the barrel; the top section of the barrel was then carefully removed. Measurements of the length of core barrel inserted into the sediment and the length of compacted sediment within the opened core were used to calculate the average sediment compaction for the sequence. The sediment was split into cylinders of 0.02 m (uncompacted) thickness. These cylinders were then weighed to 0.001 g to record the mass of wet sediment. Each sample was then air-dried until constant weight, and reweighed to 0.001 g. The change in mass of the sample following air-drying was then expressed as the sample's percentage of moisture content. The dry bulk density was determined by dividing the air-dry mass by the volume of the sample. Each sample was visually logged for colour using Munsell Soil Color Charts following the procedures of Gale and Hoare (1991), texture, relative charcoal content, macro-botanical remains, form of contacts between units, and sedimentary and/or pedological structures.

Figure 5.2 The School of Geoscience's coring pontoon on Tocal Homestead Lagoon, 1998.

Figure 5.3 Percussion coring along the A-transect using 50 mm PVC tubes, Tocal Homestead Lagoon, 1998.

The outermost material from each air-dry sample slice, that which had been in contact with the core barrel, was removed using a scalpel and stainless steel blade. This was done to minimize the risk of contamination by smearing up the inside of the core barrel during insertion. The samples from 2.00 m to 0.00 m depth (n = 100) were gently disaggregated to <2.00 mm (a single sample contained one grain >2.00 mm) using an agate mortar and pestle and the sample passed through a 2.00 mm sieve. A Retsch vibratory feeder model DR 100 and a Retsch sample divider model PT 10000 were used to split each disaggregated sample into eight sub-samples. Half of these were archived, while the remaining sub-samples were ground to a fine powder using either an agate mortar and pestle or a Retsch centrifugal ball mill type S 1000. Samples from the base of core TCA9b (3.52 m to 2.02 m, n = 75) were archived for future research.

5.3.4 Estimation of plant organic content by loss-on-ignition

The plant organic content of the sediments was estimated using the loss-on-ignition procedure of Davies (1974) (LOI_{430}). A discussion of the benefits of this procedure compared to other similar and widely employed techniques can be found in Gale and Hoare (1991). A pre-labelled porcelain crucible was weighed to 0.001 g. Approximately 5.000 g of ground sample was weighed by difference into the crucible. The crucible and its contents were then dried at 105°C for 24 hours in an oven. The crucible and sample were then desiccated until they attained room temperature, then immediately weighed to 0.001 g. The crucible plus sample were then combusted at 430°C for 24 hours in a furnace and then left to reach room temperature in a dessicator. After cooling, the crucible plus sample was weighed to 0.0001 g. The change in mass of the sample following combustion at 430°C compared with the oven-dry mass was then expressed as the percentage of material lost on ignition (LOI_{430}).

5.3.5 Mineral bulk density

The percent organic matter and moisture content data for each sample were subtracted from the total mass of each sample slice so that for the mass of dry mineral sediment per volume could be determined. Thus, dry mineral bulk density was calculated for each sample depth.

5.4 Magnetic measurements of core TCA9b

5.4.1 Setting up and calibration of magnetic equipment

All magnetic measurement equipment was set-up and calibrated following the guidelines put forward by Gale and Hoare (1991) or Walden (1999a). All measurements were made at ambient indoor room temperatures of approximately 25°C.

5.4.2 Sequence of measurements

The order in which magnetic measurements are performed is important as weak magnetic signals, such as those of natural remanent magnetisation (NRM), are destroyed by several of the measurements commonly made (e.g., saturated isothermal remanent magnetisation [SIRM]). Measurements therefore were performed in order of increasing field strength using an experimental design modified from that of Maher (1988) (Figure 5.4).

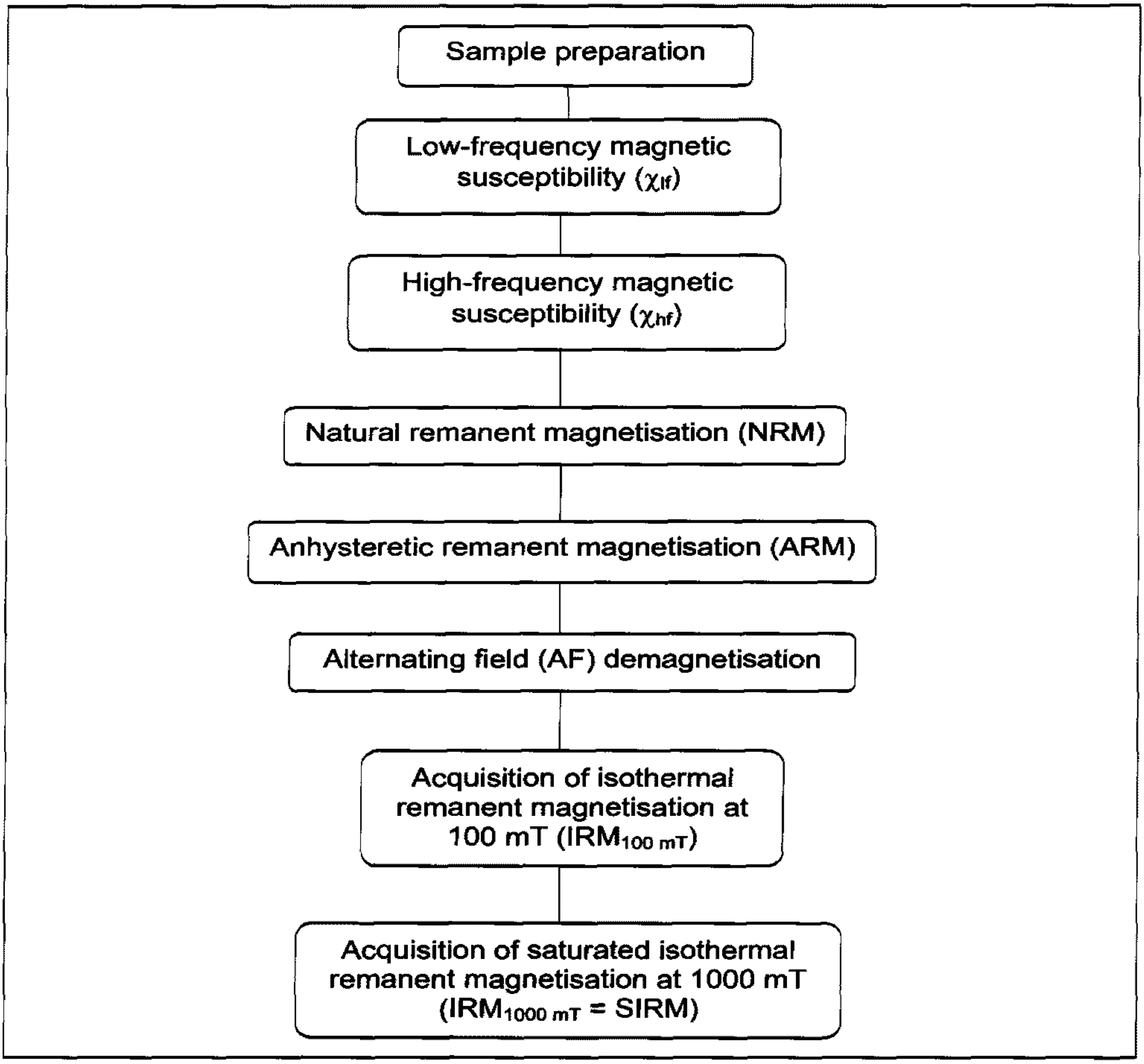


Figure 5.4 The experimental design used for the measurement of a range of magnetic properties on sediments from core TCA9b, Tocal Homestead Lagoon.

5.4.3 Sample preparation

The disaggregated sediment was tightly packed into pre-weighed polyurethane cylindrical pots of 25 cm^3 capacity and the combined mass of the pot and the sediment recorded to 0.0001 g. Magnetically screened cotton wool was used to pack the sediment tightly into the pots, thus immobilising it. Two arrows were drawn on each of the pots, one on the lid and another normal to this along the length of the cylinder. These ensured consistency of alignment of the samples within the magnetic equipment.

5.4.4 Mass-specific and frequency-dependant magnetic susceptibility

Magnetic susceptibility was measured in magnetic fields of 0.46 (low, K_{lf}) and 4.6 (high, K_{hf}) kHz respectively using a Bartington magnetic susceptibility meter model MS2 interfaced with a 36 mm internal diameter dual-frequency sensor type MS2B following the procedure outlined in Gale and Hoare (1991, pp. 221-222). The mass-specific magnetic susceptibility (χ, with units of $10^{-8}\ m^3kg^{-1}$) and frequency-dependant magnetic susceptibility (χ_{fd}, %) of each sample was determined as follows:

Eqn. 5.1 $\chi_{lf} = K_{lf} / [\ \underline{m} / \underline{V}\ ]$ Eqn. 5.2 $\chi_{fd} = 100\ [K_{lf} - K_{hf}] / K_{lf}$

Where: K_{lf} = low-field (0.46 kHz) magnetic susceptibility (10^{-5} dimensionless SI units)
K_{lf} = high-field (4.6 kHz) magnetic susceptibility (10^{-5} dimensionless SI units)
$\underline{V}$ = the bulk density of the sample relative to the density of water (m^3)
$\underline{m}$ = the mass of disaggregated < 2 mm sediment in each pot (kg)

Replicate measurements were made of susceptibility taking care to ensure consistent orientation of the samples in the MS2 meter. Thus, confidence limits of 95% were assigned to each mean data point as follows (after Gale and Hoare, 1991, pp. 49–51):

Eqn 5.3 $\underline{X} \pm t_{\alpha/2}\ [\ s / (n)^{1/2}\ ]$

Where: $\underline{X}$ = sample mean
$t_{\alpha/2}$ = t-value with an area of $\alpha/2$, where $\alpha = 0.05$
s = sample standard deviation
n = sample size

The confidence limits of both the low and high susceptibility measurements were propagated through the calculation of χ_{fd} for each sample as follows:

Eqn 5.4 $\chi_{fd}\ (\%)\ \text{error} = \{[(K_{hf}\ \text{error})^2 + (K_{lf}\ \text{error})^2] / K_{lf}\}^{1/2}$

The mass-specific magnetic susceptibility measurements made on core TCA9b were compared with those obtained from the whole-core susceptibility measurements (Figure 6.9). Except for differences in sediment compaction which are attributable to the two different coring methods utilised (Glew *et al.*, 2001) the two records were similar. This subsequently allowed data collected from core TCA9b to be extrapolated to core TCA9a, and to the remaining sediment cores from the lake.

5.4.5 Natural remanent magnetisation

The natural remanent magnetisation (NRM) of each sample was measured using a Molpsin fluxgate magnetometer. The instrument settings are given in Table 5.1. A notebook PC controlled the magnetometer, as well as resolving the net direction and magnitude of the NRM data using an in-house software program. The magnetometer was calibrated after every measurement to minimise the effect of the significant equipment drift observed (Maher *et al.*, 1999, p. 33). Calibration was maintained at ±1%; that is, if the difference between the measured and known magnetisation of the standard reference material exceeded 1% of the known value, then the magnetometer was re-calibrated. A minimum of three measurements was recorded for each sample to check that the variance for any sample remained below 5%. If necessary, measurements were repeated until this criterion was met. As measured values of NRM approached background levels, this often involved upwards of 20 measurements.

Table 5.1 Operational settings for the mineral magnetic analyses.

Magnetic measurement	**Magnetometer**		**ARM attachment**	**AF demagnetiser**	**Pulse magnetiser**
	Spin time (s)	*Attenuator*	*DC bias (mT)*	*Peak alternating field (mT)*	*Field intensity (T)*
NRM	6	1	0.04	100	-
ARM	6	1	0.04	100	-
$IRM_{100\ mT}$	6	1	-	-	100
$SIRM = IRM_{1000\ mT}$	6	10	-	-	1000

5.4.6 Anhysteretic remanent magnetisation

Anhysteretic remanent magnetisations (ARM) were induced in each sample using a Molspin AF demagnetiser with a Molspin ARM attachment. The peak alternating field used was 100 mT with a DC bias of 0.04 mT (Walden, 1999b). Samples were removed from the ARM attachment immediately after magnetisation had been induced and inserted into the magnetometer for measurement of the resulting dipole moment (M_{ARM}). Although some authors (e.g., Crockford and Willett, 1997) have suggested waiting for any viscous loss of induced magnetisation to occur before taking measurements of remanent magnetisation, the procedure of minimising the time between magnetisation and measurement was found to avoid spurious results from such losses (this was confirmed using repeat measurements on samples). The susceptibility of anhysteretic remanent magnetisation (χ_{ARM}) was determined as follows:

Eqn. 5.5 $m_{ARM} = M_{ARM} . V$

Where:
m_{ARM} = the total magnetic dipole moment of the sample due to the ARM imparted (A m^2)
M_{ARM} = the measured dipole moment (A m^{-1})
V = the volume (in m^3) of the sample which is assumed by the magnetometer during measurement. (a constant correction of 1.287 x 10^{-5}). As the samples are of unknown volume, the raw measurement must be adjusted accordingly (Snowball, 1999, p. 94).

ARM (in units of A m^2kg^{-1}) was calculated thus:

Eqn. 5.6 $ARM = m_{ARM} / \underline{m}$

Where:
$\underline{m}$ = the mass of sample used in the measurement (kg).

Most researchers have chosen to express ARM as χ_{ARM} (the susceptibility of ARM) (e.g., Snowball, 1999). ARM measurements were converted using the following relationship:

Eqn. 5.7 $\chi_{ARM} = ARM / H$

Where:
χ_{arm} = the susceptibility of ARM (m^3kg^{-1})
H = the strength of the biasing field used in inducing the magnetisation (Am^{-1}).

5.4.7 Isothermal and saturated isothermal remanent magnetisation

Some authors employ a series of increasing and/or decreasing field strengths to study magnetic behaviour in sediments. However, the large number of samples generated by this study and the vast number of remanent magnetisation measurements that would be required as a consequence meant that only two field strengths could be efficiently imparted. Many researchers have found the IRM imparted at 100 mT ($IRM_{100\ mT}$) to be of great value in investigating the magnetic mineralogy of sediments (e.g., Dearing, 1999b; Eriksson and Sandgren, 1999; Gedye *et al.*, 2000; Winter *et al.*, 2001). Walden (1999a) demonstrated that IRM may be used to approximate the total concentration of remanence carrying ferrimagnetic minerals in a sample. Thus, superparamagnetic grains that are too small to hold a stable magnetic moment, or paramagnetic minerals that do not retain a magnetisation do not contribute to IRM. Most ferrimagnetic minerals become saturated in fields of *c.* 100 mT. The value of $IRM_{100\ mT}$ is thus approximately proportional to the concentration of magnetically 'soft' ferrimagnetic minerals (magnetite, maghemite) in a sample (Thompson and Oldfield, 1986; Turner, 1997). Measurements of $IRM_{100\ mT}$ will not include any contribution from canted antiferromagnetic minerals that typically do not become saturated in fields below 1 T.

The second field strength employed is one maximising magnetisation in the samples. As the saturation remanence values of pure magnetic minerals are known (Thompson and Oldfield, 1986), measuring saturating remanence is useful in discriminating magnetic mineraology. The most common saturating field strength employed is 1 T, which not uncoincidentally is the maximum field strength that the standard Molspin pulse magnetiser can generate.

Isothermal remanent magnetisations (IRM) were induced in each sample using a Molspin pulse magnetiser to generate a field of 100 mT. These IRMs were measured using the Molspin magnetometer. As with the ARM measurements, the resulting dipole moments (M_{IRM}) were measured immediately after the magnetisation process was complete. Magnetically 'hard' canted antiferromagnetic minerals such as haematite and goethite are very unlikely to be magnetised at 100 mT, even at fine grain sizes. This sequence was then repeated using a field strength of 1000 mT. When a 1000 mT field is imparted to sediments, all ferrimagnetic and some canted antiferromagnetic minerals will reach a maximum dipole moment whereby further increases in the applied field intensity have little or no effect, at which point the sediment possess a saturated isothermal remanent magnetisation (SIRM) (Thompson and Oldfield, 1986). However, some antiferromagnetic minerals such as goethite and haematite will not be saturated in this field.

The mass-specific IRM grown in a 100 mT field ($IRM_{100\ mT}$) was calculated as follows:

Eqn. 5.8 $$m_{IRM} = M_{IRM} \,.\, V$$

Where:

m_{IRM} = the total magnetic dipole moment generated by the IRM imparted (A m^2).

M_{IRM} = the measured dipole moment (A m^{-1})

V = the volume (in m^3) of the sample which is assumed by the magnetometer during measurement (a constant correction of 1.287×10^{-5}). As the samples are of unknown volume, the raw measurement must be adjusted accordingly (Snowball, 1999, p. 94).

Eqn. 5.9 $IRM_{100\,mT} = m_{IRM} / \underline{m}$ $(A\ m^2\ kg^{-1})$

Where:
$\underline{m}$ = the mass of sample used in the measurement (kg).

The mass-specific SIRM was calculated as follows:

Eqn. 5.10 $m_{SIRM} = M_{SIRM}\ .\ V$

Where:
m_{SIRM} = the total magnetic dipole moment generated by the IRM at 1 T $(A\ m^2)$
M_{IRM} = the measured dipole moment $(A\ m^{-1})$
V = the volume (m^3) of the sample which is assumed by the magnetometer during measurement (a constant correction of 1.287×10^{-5}).

Eqn. 5.11 $SIRM = IRM_{1000\,mT} = m_{SIRM} / \underline{m}$ $(A\ m^2 kg^{-1})$

Where:
$\underline{m}$ = the mass of sample used in the measurement (in kg).

5.4.8 Magnetic ratios

A variety of ratios of magnetic measurements has been used in the mineral magnetic literature to characterise sediment magnetic mineralogy. Several such ratios have been calculated for the Tocal TCA9b data using the measurements obtained above (Table 5.2).

Table 5.2 Magnetic ratios commonly used in the analysis of lake sediments.

Magnetic ratio	Reference
ARM / χ_{lf}, χ_{ARM} / χ_{lf}	Oldfield *et al.* (1985), Oldfield and Yu (1994)
S-Ratio, $S = IRM_{100\,mT} / SIRM$	Stober and Thompson (1979), Maher *et al.* (1999)
High-field remanent magnetisation ($HIRM = SIRM - IRM_{100\,mT}$)	Maher *et al.* (1999)
$SIRM / \chi_{lf}$	Oldfield *et al.* (1985), Heider *et al.* (1996), Snowball (1999)
χ_{fd} / χ_{ARM}	Van der Post *et al.* (1997), Oldfield (1999)
$\chi_{ARM} / SIRM$	Oldfield *et al.* (1985), Maher (1998), Oldfield (1999)

5.5 Palaeomagnetism measurements

Whole-core palaeomagnetic measurements were undertaken on two cores (TCA9c and TCA9d) extracted from Tocal Homestead Lagoon using 50 mm diameter PVC barrels. During coring, no significant tilting of the core barrels from the vertical plane was noted. The core barrels were trimmed to length in the field, then sealed and transported vertically to the laboratory, where they were stored vertically prior to their magnetic measurement.

Each core was inserted into a 2G Enterprises cryogenic magnetometer following the removal of the unit's automatic AF demagnetiser unit. Stepwise demagnetization of the sediments was not undertaken in this study. The recent age of the sediments meant that any noise that may be superimposed on the secular magnetic remanence signal due to magnetic overprinting from the superimposition of more recent magnetic fields is likely to be minimal (Thompson and Oldfield, 1986; Turner and Lillis, 1994). Indeed, it is thought that the magnetic cleaning of recent sediments may degrade the natural remanence record (Thompson and Oldfield, 1986, p. 168). In any case, no equipment was available to demagnetise the sediments on a whole-core basis; that is, without undertaking the protracted procedure of subdividing each core into a series of orientated sub-samples.

Measurements of natural remanent magnetisation (NRM) in three perpendicular axis were made initially at 20 mm intervals down the core, and then subsequently at 10 mm intervals. Replicate measurements were made on each core.

The bulk density of the cores was determined via ^{137}Cs gamma ray attenuation scanning of the sediment-filled cores using a GeoTek Multi-Sensor Core Logger (MSCL). Logs of whole-core magnetic susceptibility using the Bartington MS2C sensor of the GeoTek MSCL were recorded at 10 mm (compacted depth) intervals.

5.6 Inorganic chemical analysis

5.6.1 Total inorganic chemistry by X-ray fluorescence spectroscopy

One sub-sample from each sample depth was powdered using an agate mortar and pestle and archived in plastic sample bags. In 1999, eight catchment soil samples and those samples from 2.00 m to 0.00 m in the core (n = 108) were analysed for major, and selected minor, elemental concentrations by X-ray fluorescence spectroscopy (XRF). Sample preparation followed a modified version of Norrish and Hutton's (1969) procedure. For each sample, 0.8400 (±0.0004) g of air-dried powdered sediment were mixed with 4.5000 (±0.0004) g of lithium metaborate flux (Norrish Formula) and approximately 0.06 g of ammonium nitrate. The mixture was then carefully transferred into a platinum-gold crucible and melted at 1100°C for approximately 15 minutes. Using tongs, gauntlets and a face shield, the crucibles were removed from the furnace and the resulting melt carefully poured into a glass-disc press. The press was manually operated and produced discs of approximately 40 mm diameter and 1 mm thickness.

After cooling, the discs were labelled and inserted sequentially into the sample changer of a Phillips PW2400 XRF Spectrometer. Samples were analysed for the concentrations of the following elements as oxides: Fe, Na, Mg, Al, Si, P, S, K, Ca, Ti, Ba, Ce, Cr, Mn, Ni, Cu, Zn, As, Pb, Th, Rb, U, Sr, Y and Zr. The amount of material lost during the high temperature combustion was determined by weighing several grams of each sub-sample into pre-weighed porcelain crucibles and heating them to 1000°C for one hour. The recorded loss of mass was expressed as a percentage of the original sub-sample mass (LOI_{1000} %). All results are standardised to the unvaporised component of the sample by mass.

5.6.2 Minor elemental chemistry by inductively coupled plasma-atomic emission spectroscopy

Further chemical analysis was undertaken using a Perkin Elmer Inductively Coupled Plasma-Atomic Emission Spectrometer (ICP-AES). This was done to obtain higher-resolution measurements of the minor element content of the sediments. A detailed description of the ICP technique is found in Walsh and Howie (1980) and Thompson and Walsh (1983). ICP-AES provides much lower detection limits than XRF, as well as providing improved measurements of lighter elements in the sediment matrix (e.g., Ba, Ca, S, P). Approximately 0.5000 g of powdered, air-dried sediment was added to reflux tubes and 9 cm^3 of concentrated analytical grade nitric acid slowly added. The reflux tubes were heated at 95°C for three hours on a digestion block, and then 1 cm^3 of concentrated analytical grade hydrochloric acid added. Samples were then refluxed at 110°C for one hour. Following cooling, the leachates were filtered using Whatman No. 40 (or equivalent) filter paper into pre-cleaned 100 ml conical flasks using ultra pure water and made up to the mark. Each solution was thoroughly mixed and a 10 cm^3 aliquot transferred into pre-labelled, pre-cleaned vials for analysis. The concentrations of Al, Ba, Ca, Cd, Cr, Cu, Fe, K, Mg, Mn, Na, P, Pb, S and Zn were determined simultaneously for 106 samples (99 sediment samples [there was insufficient material available to analyse one sample from TCA9b, 1.10–1.08 m], one replicate sample and six procedural blanks) by ICP-AES.

5.7 Dating core TCA9b

5.7.1 Lead-210 dating: an introduction

Detailed discussions of ^{210}Pb dating and the procedures involved have been published elsewhere (Appleby and Oldfield, 1992). Therefore only a brief introduction to the range of techniques used in ^{210}Pb dating will be presented here. Lead-210 dating requires the measurement of both ^{210}Pb and ^{236}Ra activities. Traditionally, the measurement of ^{210}Pb activity is made by alpha-counting the emissions of its granddaughter isotope, ^{210}Po. Less frequently, the emissions of its daughter isotope ^{210}Bi have been used to measure ^{210}Pb activity (Krishnaswamy *et al.*, 1971). It is only since the mid-1980s that direct counting of ^{210}Pb activity has been successfully achieved by gamma-counting following earlier and less successful attempts by Bergerioux *et al.* (1980) and others.

The gamma-counting procedure has the benefit of detecting a range of other gamma-emitting isotopes simultaneously (e.g., ^{226}Ra [via ^{214}Pb], ^{137}Cs and ^{241}Am), as well as being non-destructive. Disadvantages of using gamma-spectrometry include the much greater costs involved in obtaining counting precisions comparable with those of alpha spectrometry, due to the large investment in capital and the larger sample sizes required (approximately 50 g for gamma-spectrometry, while alpha-spectrometry on extracted ^{210}Po samples requires only approximately 5 g of sediment).

Measurements of ^{226}Ra activity are made by either alpha- or gamma-spectrometry and are often conducted separately from the measurements of ^{210}Pb. Due to the time and cost expenses involved, many studies in the past made ^{226}Ra measurements on only a selection of samples from a sequence, with the remaining values obtained by interpolation. Other studies have simply inferred a mean level of ^{226}Ra activity based on the total ^{210}Pb activity of sediments at the base of cores. These sediments are thought to have no unsupported ^{210}Pb activity and to be in equilibrium with the parent isotope ^{226}Ra. As Ivanovich and Harmon (1992) have pointed out, such assumptions do not account for variations in sedimentation rates, changes in sediment sources or alterations in organic productivity. Most recent studies on ^{210}Pb chronologies have made individual measurements of ^{226}Ra for every ^{210}Pb determination, allowing the variation in ^{226}Ra activity to be accounted for (e.g., Gale *et al.*, 1995).

It should be noted that several other methods exist that may be used to measure radionuclide activity. These include neutron activation analysis, mass (and accelerated mass) spectrometry, and fission and alpha track methods. These methods are generally only suitable for the measurement of the heavy radionuclides in the U-Th decay series. Although mass spectrometrical techniques are commonly used to examine lead isotopes in environmental materials, the extremely low detection limits required for ^{210}Pb dating make them unsuitable for this purpose.

5.7.2 Lead-210 dating of core TCA9b

The ^{210}Pb activity of 44 samples from the top 1.26 m of core TCA9b was determined by alpha spectrometry. All reagents used were analytical grade or superior. All reagent water used was filtered by reverse osmosis or better.

Between 3.0000 and 5.0000 g of air-dried powdered sediment were weighed into a 150 cm^3 beaker. Approximately 600 Bq of ^{133}Ba and ^{209}Po tracers were added by weighing 0.10 g of tracer stock of known concentration into each sample. The activities of the tracer solutions were as follows: ^{133}Ba (6 008 400 ± 9360 Bq g^{-1}, as of 17 May 1993) and ^{209}Po (6600 ± 60 Bq g^{-1} as of 4 March 1998). Twenty-five cubic centimetres of concentrated nitric acid were added in 5 cm^3 increments, and a drop of *n*-octanol added to prevent excess foaming. Samples were then evaporated to dryness on a water bath. Each beaker was removed from the water bath and *c.* 25 cm^3 of 10% hydrogen peroxide slowly added to assist in the decomposition of organic matter.

Once the reaction had stabilised (>15 min), samples were placed on the water bath and heated until any further effervescence had ceased. Twenty-five cubic centimetres of concentrated hydrochloric acid were added to each sample and each beaker covered with a watch glass to produce an azeotrope. Samples were refluxed for 6 hour on the water bath and the contents quantitatively transferred to a 50 cm^3 centrifuge tube, made up to volume with 6 M hydrochloric acid and centrifuged for 2 minutes at 5100 rpm. The supernatants were then transferred to 250 cm^3 separating funnels and excess iron extracted with diethyl ether. The extracts were evaporated to dryness on a water bath.

The samples were rehydrated by the addition of 50 cm^3 of 0.1 M hydrochloric acid and 30 cm^3 of reagent water. Any trace iron and chromium in the solution was complexed by the addition of 100 mm^3 1.0 M citric acid solution, and the pH of the solution adjusted to 1.5.

The extract was reduced to its chloride form by the addition of 1 g of hydroxylammonium chloride. Polonium was autodeposited onto silver discs for a minimum of 4 hours. The discs were then placed in plastic petri-dishes and the ^{210}Po activity measured by alpha spectrometry (using passivated implanted planar silicon detectors).

The remaining solution was made up to 800 cm^3 with distilled water, and 20 cm^3 of concentrated sulphuric acid and 100 cm^3 of 20% w:v sodium sulphate were added. A hundred milligrams of lead in 0.1 M nitric acid were slowly added and the radium-barium-lead sulphate precipitate allowed to settle overnight. The supernatant was removed, and the precipitate was separated by centrifugation and dissolved in a 0.2 M alkaline diethylenetriamine-pentaacetic acid (DTPA) solution.

Radium-barium sulphate was selectively precipitated at pH 4.5 by the simultaneous addition of 2 cm^3 of 1:1 v:v acetic acid/water and 1 cm^3 of barium sulphate seeding solution. The colloidal radium-barium sulphate precipitate was collected on a smooth-surfaced Millipore 'VV' membrane filter in a lock-seal filter apparatus. The ^{226}Ra peaks at 4.602 and 4.784 MeV in the alpha-spectrum were measured by alpha spectrometry of the filter source. The alpha spectrometers used were an EG&G Ortec model Octeté PC and a Tennelec model TC256. The alpha spectrometers were energy and efficiency calibrated with an ^{243}Am standard source prepared by electrodeposition of an aliquot of ^{243}Am stock solution (127 200 ± 1140 Bq g^{-1}) and a ^{226}Ra source co-precipitated with $BaSO_4$ (82 380 ± 900 Bq g^{-1}).

Recovery was estimated by measurement by gamma spectrometry of the ^{133}Ba peak at 356 keV in the gamma-spectrum for a minimum of 10 min using an EG&G Ortec model GMX-20190 N-type high purity germanium (HPGe) gamma-ray spectrometer. The HPGe spectrometer was energy calibrated using a National Institute of Standards and Technology (NIST) traceable $^{154}Eu/^{155}Eu/^{125}Sb$ multi-nuclide standard. The tracer yield was determined by comparing the ^{133}Ba activity in each sample with the known ^{133}Ba activity of a standard with the same geometry.

The activity of the supported ^{210}Pb in the sediments was assumed to be in equilibrium with that of ^{226}Ra, while total ^{210}Pb activity was taken to equal ^{210}Po activity. The activity of the unsupported ^{210}Pb was therefore determined by subtraction of the supported from the total ^{210}Pb activity. The ^{209}Po and ^{210}Po activities where measured using the 4.88 keV and 5.30 keV peaks respectively. The ^{226}Ra activity was measured using the 4.79 keV peak. Uncertainties quoted are 1σ counting errors propagated from the measurement of ^{209}Po and ^{210}Po activities.

5.7.3 Caesium-137 dating of core TCA9b

Sub-samples from the upper section of core TCA9b were selected for measurement of their ^{137}Cs activity. By examination of the downcore pattern of ^{137}Cs in the upper part of the core, an independent chronology may be developed for the last 40–50 years based upon the first appearance of the anthropogenic isotope in the sediment (Brunskill *et al.*, 2002a).

The ^{137}Cs activity of samples at the following depths from core TCA9b were determined by Compton suppression gamma spectrometry; 0.04–0.02 m, 0.18–0.16 m, 0.26–0.24 m, 0.36–0.34 m, 0.40–0.38 m, 0.44–0.42 m, 0.46–0.44 m, 0.48–0.46 m, 0.50–0.48 m, 0.52–0.50 m, 0.54–0.52 m, 0.56–0.54 m, 0.58–0.56 m, 0.60–0.58 m, 0.62–0.60 m, 0.64–0.62 m and 0.80–0.82 m. The Canberra Compton suppression detector system comprises an active NaI(T1) suppression annulus, a NaI(T1) plug detector and a reverse electrode germanium (REGe) detector all housed within an inert lead shield. Air-dried and ground sub-samples were sealed into 55 mm petri dishes and counted for between 30 and 140 hours depending on sample mass and ^{137}Cs activity. The ^{137}Cs activity of each sample was determined using the 662 keV peak after subtraction of the ^{214}Bi peak interference. Uncertainties quoted are 1σ counting errors. The detector system energy calibration was carried out using a NIST traceable $^{154}Eu/^{155}Eu/^{125}Sb$ multi-nuclide standard source and the detector system efficiency calibration was determined by counting the International Atomic Energy Agency Soil-6 reference material.

5.7.4 Carbon-14 dating of core TCA9b

Five radiocarbon dates have been performed on samples extracted from the top 2.00 m of core TCA9b. Two pieces of charcoal were initially extracted from positions of interest in the stratigraphy of core TCA9b (1.32–1.30 m and 1.28–1.26 m) in 1998 and were submitted to the University of Waikato, in collaboration with the Rafter Radiocarbon Laboratory of the Institute of Geological and Nuclear Sciences (IGNS) at Lower Hutt, New Zealand, for radiocarbon dating. A single charcoal fragment was submitted to the Australian Nuclear Science Technology Association (ANSTO) Radiocarbon Dating Laboratory for Accelerator Mass Spectrometry (AMS) ^{14}C dating. Finally, two samples of microscopic organic material extracted from core TCA9b (1.28–1.26 m and 1.26–1.24 m) were submitted to Beta Analytic Inc., USA for AMS ^{14}C dating. All ^{14}C measurements and sample pre-treatments were carried out by specialists at the three laboratories listed above.

Charcoal samples were extracted from the dried sediment samples using a pre-cleaned scalpel and tweezers. Microscopic organic samples were extracted following the procedure of Regnell (1992). Particular attention was taken to ensure that no contamination of the charcoal samples took place, either from adjacent sediment samples or during extractions.

The principles of both conventional and AMS ^{14}C dating are given in Geyh and Schleicher (1990). A brief summary of the techniques used in the analysis of the charcoal from Tocal is presented here instead. Submitted samples were ignited to produce CO_2 gas, which is captured in a glass ampoule for measurement in an accelerator mass spectrometer (Taylor, 1987). The gaseous sample is placed in the spectrometers ion source, ionised by bombardment with caesium ions and focused into a high-velocity ion beam. The first magnet in the mass spectrometer attracts ions of a defined atomic mass (in this instance 14). The ions $^{12}CH_2$ and ^{13}CH, and a very few ^{14}C ions are then directed into the accelerator in order to separate the ^{14}C ions from the mixture. A second magnet then selects ions with the momentum expected of ^{14}C ions into a final detector for counting (Gillespie, 1984).

5.8 Tocal catchment palaeosol

5.8.1 Sampling

Webber's Cottage is a small building located on the hill immediately west of Tocal Homestead Lagoon. The cottage was constructed in the early 1820s and is considered to be the oldest remaining structure in the catchment. A suspended-floor construction in the main room has allowed the underlying soils to remain undisturbed since construction of the cottage. A trapdoor in the room's corner provided limited access to the soil and two cores (TCWC1, and TCWC2) were extracted by percussion coring using PVC pipe. The soil was found to be weakly structured and friable, with rare rounded gravels that made coring problematic. The first attempt at sampling this profile used 101 mm diameter pipe. This was found to be too wide to retain the soil during removal, and only the top *c.* 60% was obtained, the remainder of the soil falling out of the core barrel during extraction. A 50 mm diameter core (TCWC1) was then used and 0.515 m of material removed. A duplicate 50 mm core (TCWC2) was also taken, which sampled the top 0.321 m of the soil profile. The cores were labelled, any remaining air-space packed with paper and the top and bottom of each core tightly sealed using thick plastic and tape. The cores were then transported vertically to the laboratory, where they were stored vertically prior to analysis.

5.8.2 Soil core logging, physical and magnetic measurements

Volume-specific magnetic susceptibility was measured on each core (following the procedure detailed in section 5.3.1) with the loop sensor aligned horizontally on a purpose-built framework that enabled cores to be measured while remaining vertical. This avoided any disturbance of the soil profile prior to subsampling. Repeat measurements were performed for each core. Using the depth of insertion recorded for each core, the average compaction of the soil due to coring was determined. Based on these data, and the depth of soil sampled, core TCWC1 was selected for more detailed measurements. The core barrel was opened by making vertical cuts along opposite sides of the barrel. The top section of the core barrel was then carefully removed and the soil profile logged for visual properties. The soil was split into cylinders of 0.02 m (uncompacted) thickness. Moisture content and bulk density were measured following the procedure used for the lake sediment samples (section 5.3). Sub-samples were disaggregated, sieved at 2.00 mm, split, powdered and loss-on-ignition (LOI_{430} %) measurements made using those procedures employed for the lake sediment samples (section 5.3).

5.8.3 Separation of the mud and sand fractions

Air-dried and ground subsamples of the Tocal palaeosol were leached for chemical analysis using a sequential extraction procedure. An experimental design was chosen that allowed the chemical contribution of the <63 μm (mud) and the >63 μm (sand) size fractions of the soil to be identified.

Air-dried and powdered soil samples were wet-sieved at 63 μm using a non-metallic sieve with nylon mesh and Millipore ultra pure water. The material retained on the sieve was transferred into 250 cm^3 beakers using excess ultra pure water. The liquid phase was evaporated off in a low temperature (38°C) oven, thus avoiding the possible vaporisation of elemental Se, As, I, Sb, Hg and Cr from the sample. The <63 μm fraction and water were retained in an oversized evaporative dish. The contents were transferred into 1000 cm^3 glass beakers using excess ultra pure water and the liquid phase also evaporated off at 38°C. The second set of sub-samples therefore contained both the mud-sized fraction and the water soluble fraction of the original bulk soil sample. It should be noted that the sand sub-sample did not include the water soluble component of the >63 μm fraction, which was incorporated in the 'mud' sample as an artefact of the experimental design. The sub-sampling procedure is shown schematically in Figure 5.5.

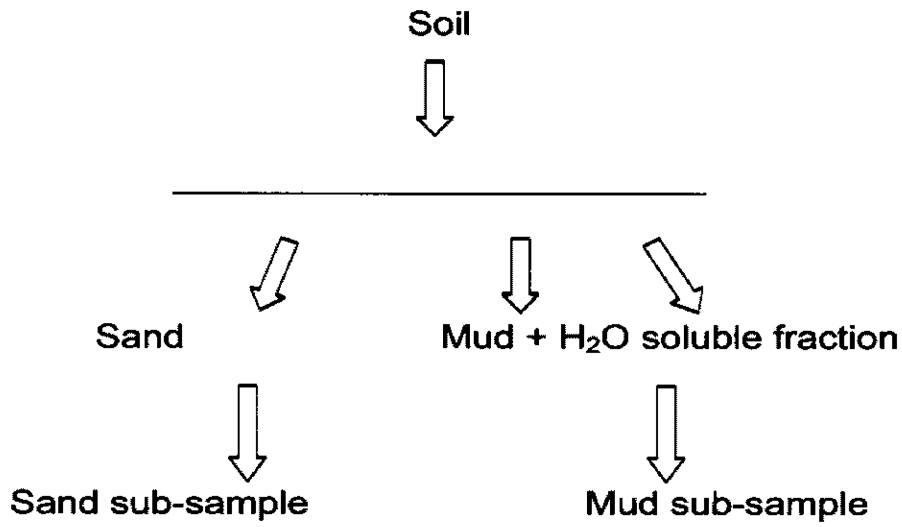


Figure 5.5 The procedure used in separating the sub-fractions of the Tocal palaeosol for chemical analysis.

5.8.4 Mud sub-sample preparation – weak acid leach

Only the mud sub-samples were selected for weak acid leaching using Millipore ultra pure water. The water soluble fraction of the sand sample is assumed to have been removed during the separation procedure and retained as part of the mud sub-sample (Figure 5.5). The oven-dried mud subsamples were allowed to equilibrate to ambient humidity for a minimum of 48 hours. Approximately 1.5000 g of the mud sub-sample was added to 50 cm^3 centrifuge tubes which had been cleaned in 5% v:v Decon90 solution and 2.5% v:v nitric acid. Twenty cm^3 of ultra pure water was added to each tube, followed by 1–2 drops of $CaCl_2$ to discourage reabsorption (Schramel *et al.*, 2000). The centrifuge tubes were shaken for 6.0 hours on a rotating shaker and then centrifuged for 10 minutes at 5000 rpm. The supernatant was passed through Whatman No. 40 filter paper into pre-cleaned 100 cm^3 volumetric flasks and made up to the mark using ultra pure water. For every 20 samples processed, one sample duplicate was generated. Two blank solutions were prepared as above. All solutions were then stored temporarily at approximately 5°C.

5.8.5 Mud and sand sub-sample preparation – *Aqua Regia* leach

Approximately 1.0000 g of material from each of the mud and sand sub-samples was added to labelled digestion tubes. While in a ventilated fume cabinet, 10 cm^3 of ultra pure water was added to each tube followed by the careful addition of 10 cm^3 of *Aqua Regia* (3:1 v:v analytical grade concentrated nitric and hydrochloric acids). Watch glasses were placed on top of each of the digestion tubes, and the reaction accelerated by placing the tubes in a heating block for 4 hour at approximately 100°C. During the heating, the labels on the tubes began to vanish, and each tube was relabelled at least once. The tubes were allowed to cool, and the contents filtered using Whatman No. 40 filter paper into labelled and pre-cleaned 100 cm^3 volumetric flasks using ultra pure water. Sample duplicates and procedural blanks were prepared as described in section 6.8.4. Three analytical standards were created by digesting 1.0 g of AGAL-10 (Australian Government Analytical Laboratories Reference Material – Hawkesbury River Sediment) following the same procedure used for the palaeosol samples.

5.8.6 Soil leachate analysis by ICP-AES

Each volumetric flask was stoppered and shaken vigorously to ensure that the soil solution was completely homogenised, then a 10 cm^3 aliquot of the contents was transferred into a sterile, labelled glass vial. The elemental concentration of each of the digestions was determined using a Perkin Elmer ICP-AES in the School of Chemical Engineering at The University of Sydney. The elements analysed for were Al, As, Ba, Cd, Co, Cr, Cu, Fe, Mn, Mo, Ni, Pb, V and Zn.

5.8.7 Emerson's aggregate test for structural stability

A soil's erodibility is an important factor in any study of soil loss. Soil aggregate stability is one important element in controlling soil erodibility, and its measurement may provide information on the relative ease with which individual soils may be eroded. Emerson's aggregate test (EAT) (Emerson, 1967) is a qualitative field or preliminary laboratory procedure used to assess the structural stability of aggregates, typically of between 2 to 5 mm in diameter. Other measures of aggregate stability exist, for example, that of Kemper and Rosenau (1986), but are more time consuming and laborious to execute. Emerson's test is adequate for preliminary comparison of the aggregate stabilities of present-day catchment soils at Tocal with those of the palaeosol (TCWC1). The test involves immersing air-dried aggregates in water and recording the aggregates' response. These observations are then employed to allocate a stability class ranging from 1 (least stable) to 8 (most stable) (Table 5.3). The stability classes have recently been expanded by Charman and Murphy (1991) for use on Australian soils, with the division of classes 2 and 3 into four subclasses. These modifications were not employed here as the additional sophistication of Charman and Murphy's procedure was deemed unnecessary in a qualitative, preliminary investigation of soil aggregate stability.

Table 5.3 Soil aggregate behaviour and the determination of EAT class (Emerson, 1967).

EAT class	Summary of observations of soil aggregates
1	Slaking and complete dispersion
2	Slaking and some dispersion
3	Slaking and some dispersion after remoulding
4	Slaking and no dispersion (Carbonate or Gypsum present)
5	Slaking and dispersion in shaken suspension
6	Slaking and flocculation in shaken suspension
7	No slaking and aggregates swell in water
8	No slaking and no swelling of aggregates in water

5.8.8 Soil aggregate stability

Perhaps the most widely used method of assessing soil aggregate stability is that of Kemper and Rosenau (1986). It is the standard laboratory method used by the United States Department of Agriculture (USDA) Soil Survey Laboratory (Anon., 1996). The procedure uses sieving to simulate the disruption of soil aggregates by water. Minor modifications to the procedure were made in accordance with the availability of equipment in the laboratory and soil for analysis. These changes are not considered to have had any effect on the results obtained.

A 2.0 mm brass sieve was assembled on top of a 1.0 mm brass sieve. Each sample was disaggregated by gloved hand or mortar and pestle to just pass through the 2.0 mm sieve. The material retained on the 2.0 mm sieve was removed, and the 2.0 to 1.0 mm sized material was sieved again using the 1.0 mm sieve to remove excess dust. 3.00 $\pm$ 0.05 g (or as close to 3.00 g as possible, given limitations in the amount of sample available) of the 1.0–2.0 mm component was weighed by difference into glass petri dishes. A 0.5 mm brass sieve was placed in an over-sized evaporative dish and tap water slowly added until the water level was 20 mm above the base of the sieve screen. The 3.00 g sample was distributed on the sieve screen such that no aggregates were in contact with one another, and each sample was left overnight (*c.* 12 hours).

The samples were then agitated by raising and lowering the sieve through the 20 mm of water at a rate of 20 times in 40 seconds. With each 'stroke', the sieve was raised sufficiently for the sieve to drain but not raised so high that air could enter between the base of the sieve and the water surface.

The sieve was removed from the dish and placed on a sheet of aluminium foil larger than the diameter of the sieve in a drying oven for 3 hours at 105°C. Aggregates that continued to disintegrate during the drying process were retained on the aluminium foil. The sieve and retained sample, and the foil sheet were removed and allowed to cool until they had reached ambient room temperature and humidity. The aggregates retained on the sieve were removed with a thick bristle brush into a tared petri dish and weighed using an electronic balance to 0.01g. Aggregate stability was then reported as the percentage of aggregates (%) (in the 0.5 to 2.0 mm size range) retained after wet sieving by the following equation:

Eqn. 5.11 $$\text{Aggregate Stability} = [\, M_a / M_t \,] \times 100$$

Where:
M_a = Mass of soil aggregates retained on 0.5 mm sieve (g)
M_t = Mass of original sample (g)

The alternative steps to be taken if sand >0.5 mm is found in the sample were not necessary in the case of the Tocal palaeosol (Anon, 1996, p. 196).

5.9 Tocal catchment soils

5.9.1 Sampling

A soil map of the Tocal property developed by Agriculture NSW, incorporated into the Tocal GIS (section 5.1), was used to choose locations from which to sample soils in the Tocal catchment. Eleven soil groups were identified within the catchment, with six instances of one group (group 23) recorded (Figure 5.6). Thirteen soil groups were selected for sampling. Sample locations were selected using a random co-ordinate generator algorithm. For each soil group, five random pairs of co-ordinates were generated. This allowed any pair of co-ordinates to be discounted if they represented locations that were impossible to sample (such as on a road or in the lake). Sample locations were located using hand-held GPS units. At each location a composite sample was collected from four points arranged in a cruciform pattern at a distance of 0.5 m from the sample site. A 90 mm PVC pipe and percussion coring equipment (section 5.2.1) was used to collect 0.15 m deep samples at each corner of the four points. The four samples were removed from the PVC pipe in the field and were combined in a plastic container, which was then sealed and transported back to the laboratory.

5.9.2 Sample preparation

Each bulk soil sample was homogenised, air-dried, disaggregated, sieved at 2.0 mm, and split following the procedures detailed in section 5.3.3. One set of sub-samples (T-series) was ground to a powder using a porcelain mortar and pestle, while a shatter-box with a tungsten carbide mortar and hammers was used to powder a second set of sub-samples (TS-series) in preparation for chemical digestion and analysis. By doing this, it was hoped to be able to assess any impact that grinding methods might have on chemical recovery.

5.9.3 Acid leach

Approximately 0.5000 g of soil was added to labelled digestion tubes. While in a ventilated fume cabinet, 9 cm^3 of concentrated analytical grade nitric acid was added to each digestion tube, which was then placed in a heating block for 4 hours at approximately 100°C. The tubes were removed and 1 cm^3 of concentrated analytical grade hydrochloric acid added. The tubes were replaced in the heating block for a further 2 h at approximately 100°C. The tubes were allowed to cool, and the contents filtered using Whatman No. 40 filter paper into pre-cleaned 100 cm^3 volumetric flasks using ultra pure water. Sample duplicates and procedural blanks were prepared as described in 6.9.2. One analytical standard was created by digesting 1.0 g of AGAL-10 (Australian Government Analytical Laboratories Reference Material – Hawkesbury River Sediment) following the same procedure used for the soil samples.

5.9.4 Soil acid leach analysis by ICP-AES

Each volumetric flask was stoppered and shaken vigorously to ensure that the soil extract was completely homogenised, then a 10 cm^3 aliquot of the contents was transferred into a sterile, labelled vial. The elemental concentration of each of the digestions was determined using a Perkin Elmer ICP-AES in the School of Chemical Engineering at The University of Sydney. The elements analysed for were Al, Ba, Ca, Co, Cr, Cu, Fe, K, Mg, Mn, Na, Ni, P, Pb, Se, Sr, V, Zn. The original aliquots were too concentrated for the determination of Al, Fe and (for one sample) Mn. These were diluted and re-measured by the ICP-AES. One sample (TS4w) was selected for duplicate extraction and analysis. To check for analytical precision, three samples (T18g, T23e, TS18g) and a procedural blank were re-analysed by ICP-MS.

Figure 5.6 Tocal catchment soil map, with the watershed marked in red. Numbered, coloured regions on the map represent different soil types identified in the Tocal Homestead Lagoon catchment as part of the Agriculture New South Wales GIS (used with permission)(see Table 2.1 for key). Asterisks mark the randomly selected sample locations.

Chapter Six

Results

6.1 Lake sediment cores

6.1.1 Data confidence limits

Where possible, the data presented here incorporate all the raw laboratory measurements made for each sample. This has allowed confidence limits to be included for most data points. Unless otherwise noted, mean values are reproduced, while error bars represent 95% confidence limits of the measurements.

6.1.2 Volume-specific magnetic susceptibility

The results from the whole-core measurement of volume-specific magnetic susceptibility (VSMS) for the 21 sediment cores extracted from Tocal Homestead Lagoon are presented in Figure 6.1 (see also Appendix 1). The cores demonstrate that a similar pattern of down-core susceptibility exists across the entire basin. The lowermost sediments display quite low susceptibilities (10–20 x 10^{-5} SI units). Values increase rapidly above this background level, fluctuating through several peaks and troughs to the tops of the respective cores. Peak values differ between cores, but are generally in the range of 150–250 x 10^{-5} SI units. This pattern of peaks and troughs is repeated across the basin, with minor variations at each core location.

The consistent pattern of susceptibility from core to core is indicative of a similarly consistent pattern of sedimentation across the lake basin. The existence of common magnetostratigraphic features also allows the correlation of the sediment cores, connecting points of contemporaneous deposition (Figure 6.2). Cores from the centre of the basin, such as TCA9a, possess high-resolution susceptibility records, while marginal cores such as TCB6 record far fewer features. The cores extracted from the margins of the lake (TCA1, TCA2, TCB1, TCB2, TCB6, TCC8 and TCE9) have much shorter susceptibility records, and are more likely to have experienced subaerial weathering or gaps in deposition at periods of low lake levels (Rolph *et al.*, 1996; Gale and Haworth, 2005). The result is the development of more complex susceptibility records that are more difficult to correlate with those of other cores. Records from marginal locations in the lake are also more susceptible to distortion from reworking due to lake-level fluctuations, shallow-water wind effects (Lloyd *et al.*, 1998) or bioturbation from livestock.

6.1.3 Temporal and spatial patterns of lake sedimentation

The magnetostratigraphy established in Figure 6.2 demonstrates how lake sedimentation has varied through time and over space, across the basin of Tocal Homestead Lagoon. Rates of sedimentation within the lake are greatest along the central axis of the basin, primarily in the areas of greatest water depth such as TCA8 and TCA9. A comparison of the sediment accumulation pattern as revealed by the magnetostratigraphy with the lake's bathymetry (Figure 2.6) confirms that water depth exhibits some measure of control on sediment accumulation rates. Surprisingly, there appears to be no evidence of significant fluvial deposition in the lagoon; that is, there are neither thicker deltaic accumulations near the mouth of the stream that feeds the lake, nor decreases in sediment accumulation away from the southwest limb. This suggests that the contribution of sediment transported by fluvial processes into the lake is relatively minor, lending support to the argument that slope processes of sediment delivery may be dominant in this system.

6.2 Core TCA9b physical properties

6.2.1 Core description and stratigraphy

The visual log of core TCA9b is presented in Figures 6.3 and 6.4. All colours depicted in these figures are the Munsell soil colours recorded (detailed in section 5.3.3) converted to CMYK print colours using the Munsell Conversion Program – Version 6.5.1 computer software. The lake sediments are composed predominantly of grayish brown (10YR 5/2) to light brownish gray (10YR 6/2) silt, with a few layers containing very rare fine sand grains. A single rounded quartz grain of 3 mm diameter was found at 0.94–0.92 m. Laminations were frequently observed; these varied in thickness from 1 mm to 10 mm. Mottling of up to 70% of the sediments and possible pedogenesis was noted, typically in those sediments from the middle of the core (1.18 to 0.68 m). Most of the sediments in the core contained at least low concentrations of macroscopic plant organic matter, usually fine rootlets in growth position. Sporadic and isolated layers of highly concentrated plant remains were also noted. These consisted of well-developed *in situ* root systems. Generally the plant organic remains were well preserved, though sediments from 0.92 to 0.68 m also contained plant remains in various states of decomposition. A single, large piece of *Eucalyptus* root wood (John Ford, personal communication, 1999), approximately 13 cm long was found aligned horizontally at a depth of 0.99 m.

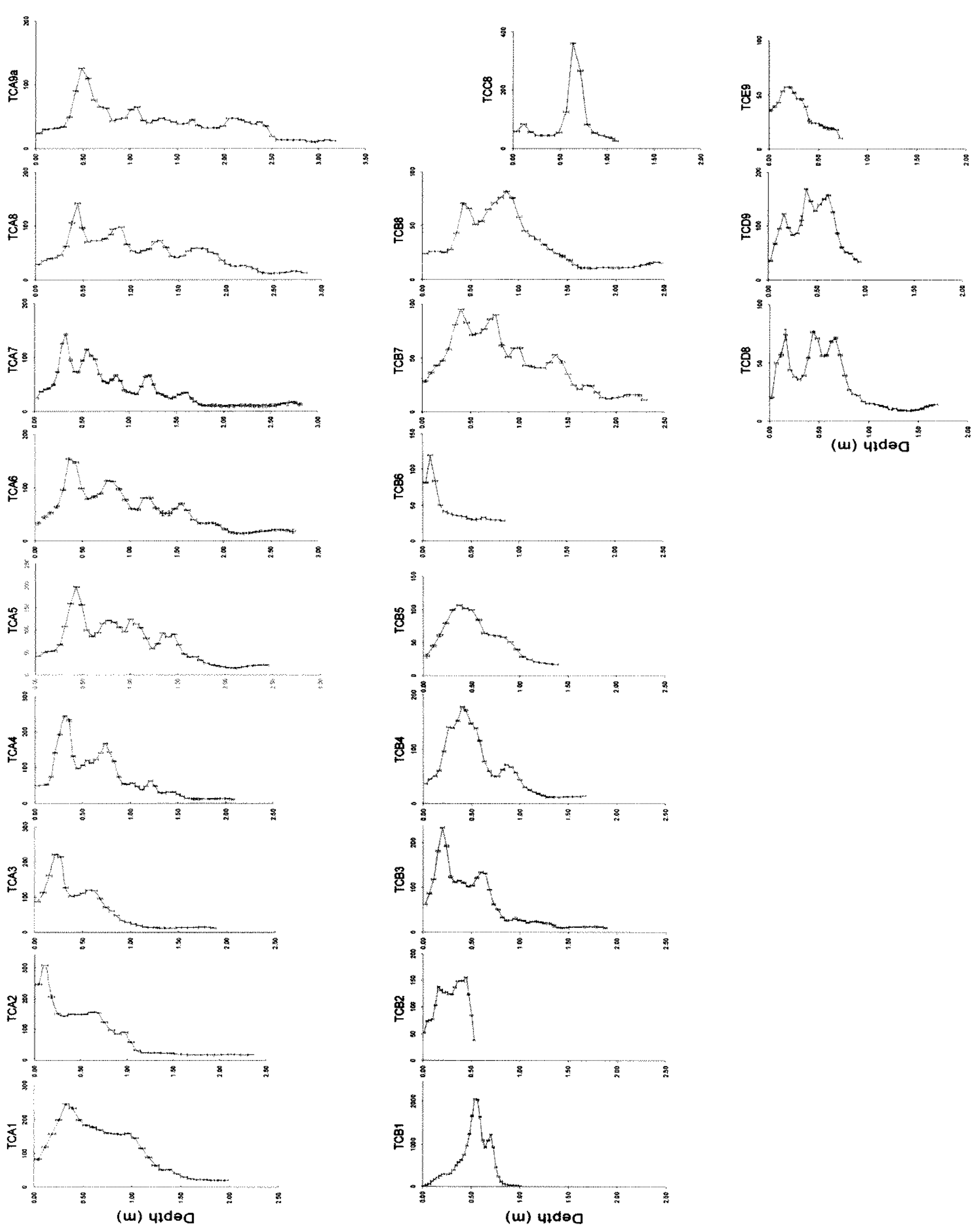
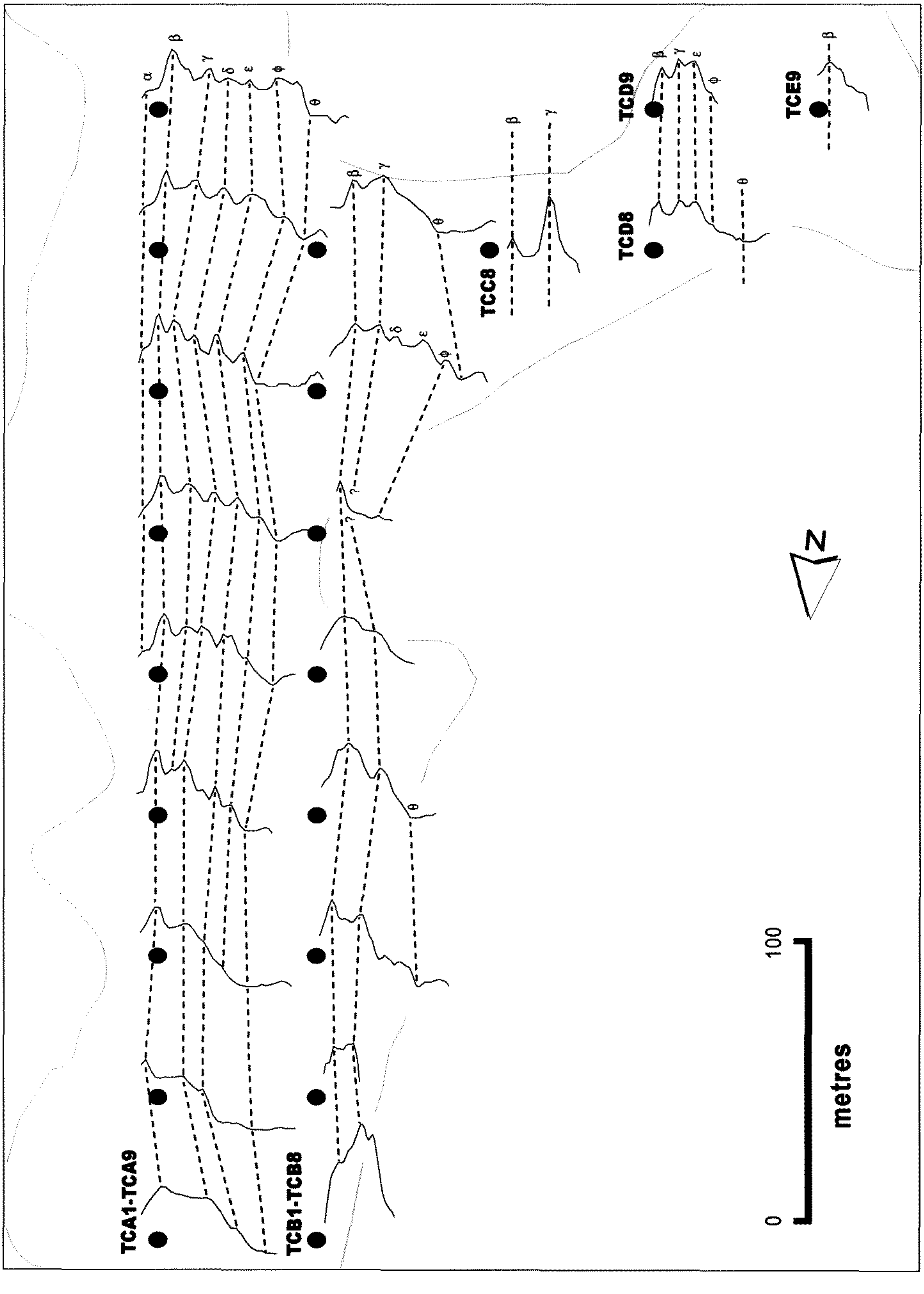


Figure 6.1 Downcore variations in whole-core magnetic susceptibility (x 10^{-5} SI units) for 21 cores extracted from Tocal Homestead Lagoon, Paterson, eastern Australia. Horizontal error bars are at 2 sigma confidence intervals, while vertical error bars span sample thicknesses. Raw data are tabled in Appendix 1.


TCA1-TCA9
TCB1-TCB8
TCC8
TCD8
TCD9
TCE9
N
0
100
metres
α
β
γ
δ
ε
φ
θ

The sediments from 2.00 m to 1.14 m contain very rare to moderate concentrations of macroscopic charcoal fragments generally of 1–3 mm long axis. Two exceptions are those sediments from 2.00–1.98 m and 1.74–1.72 m, which contained no macroscopic charcoal. The charcoal fragments were found to be evenly dispersed throughout the sediment slices. Charcoal was generally absent above 1.14 m. Three discrete layers of charcoal were observed in the remainder of the core, at 0.98–0.96 m, 0.64–0.60 m and 0.52–0.48 m. Those sediments from 0.98–0.94 m and 0.52–0.48 m contained moderate amounts of charcoal, while charcoal was rare from 0.64–0.60 m. Charcoal fragments of up to 50 mm long axis were present in the sediments from 0.64–0.60 and 0.52–0.48 m.

Macroscopic crystals of vivianite ($Fe_3(PO_4)_2.8H_2O$) were observed above 0.92 m in core TCA9b. Moderate concentrations of vivianite were recorded at 0.90–0.86 m and at 0.62–0.60 m. Vivianite was very rare to rare in the remainder of those sediments from 0.92 to 0.54 m, with no vivianite observed at 0.68–0.66 m. The sediments above 0.54 m contained no vivianite crystals. Vivianite tended to be associated with macroscopic charcoal, and to a lesser extent plant organic matter, in the sediments. Partial and even complete replacement of charcoal fragments by vivianite was observed (especially at 0.62–0.60 m).

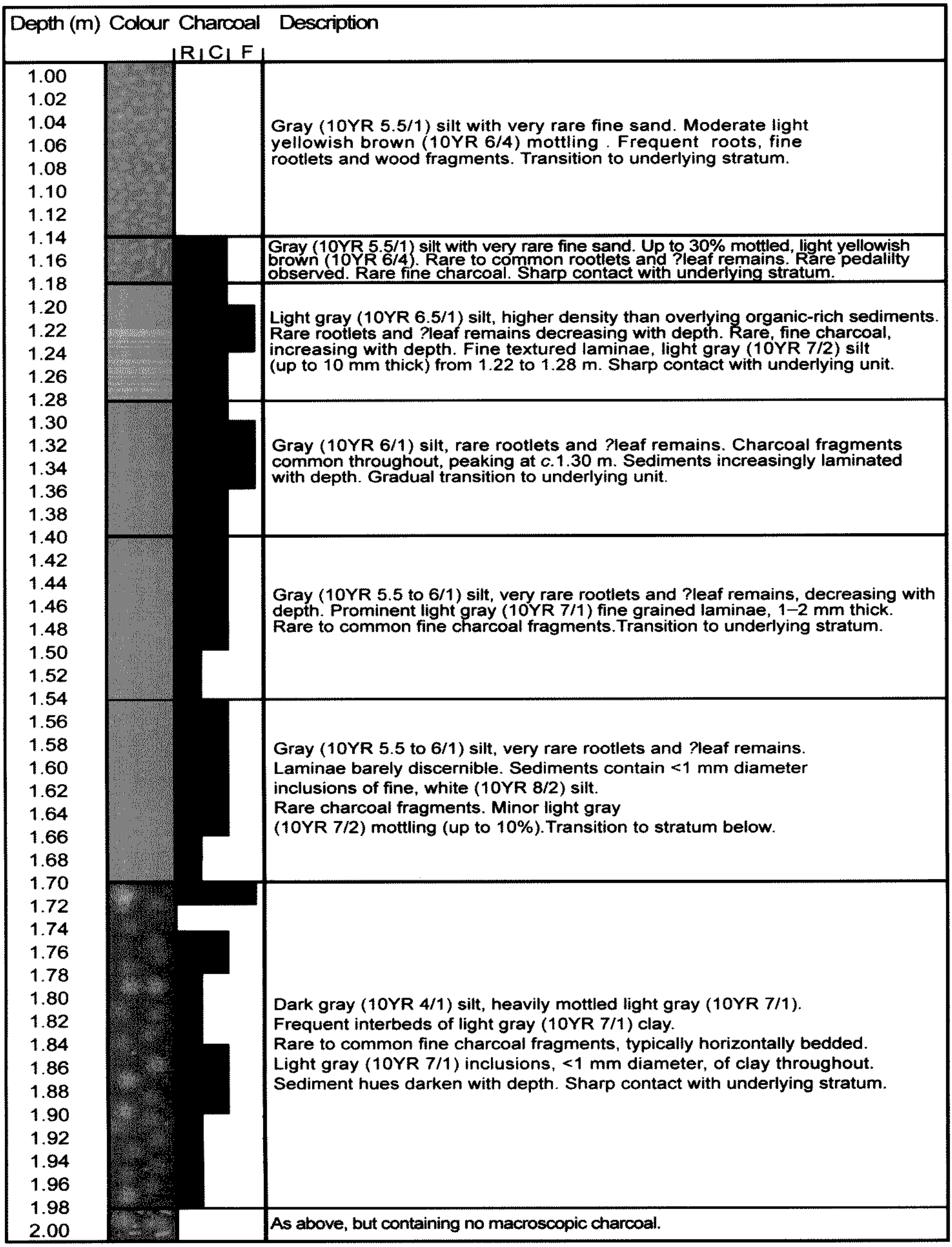


Figure 6.3 Visual log of the sediments from 2.00 to 1.00 m depth, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. The letters R, C and F denote the rare, common and frequent occurrence of macroscopic charcoal fragments.

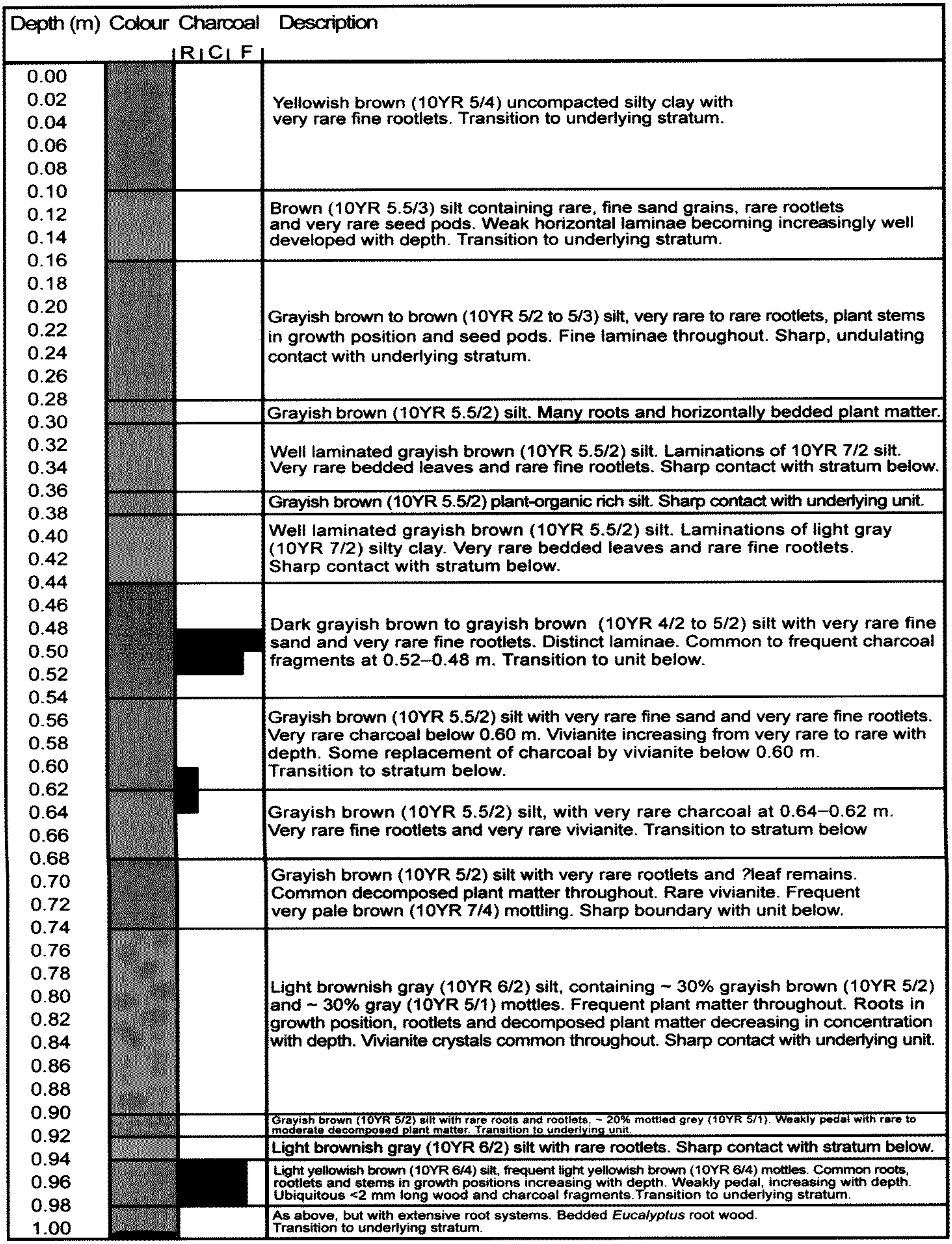


Figure 6.4 Visual log of the sediments from 1.00 to 0.00 m depth, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. The letters R, C and F denote the rare, common and frequent occurrence of macroscopic charcoal fragments.

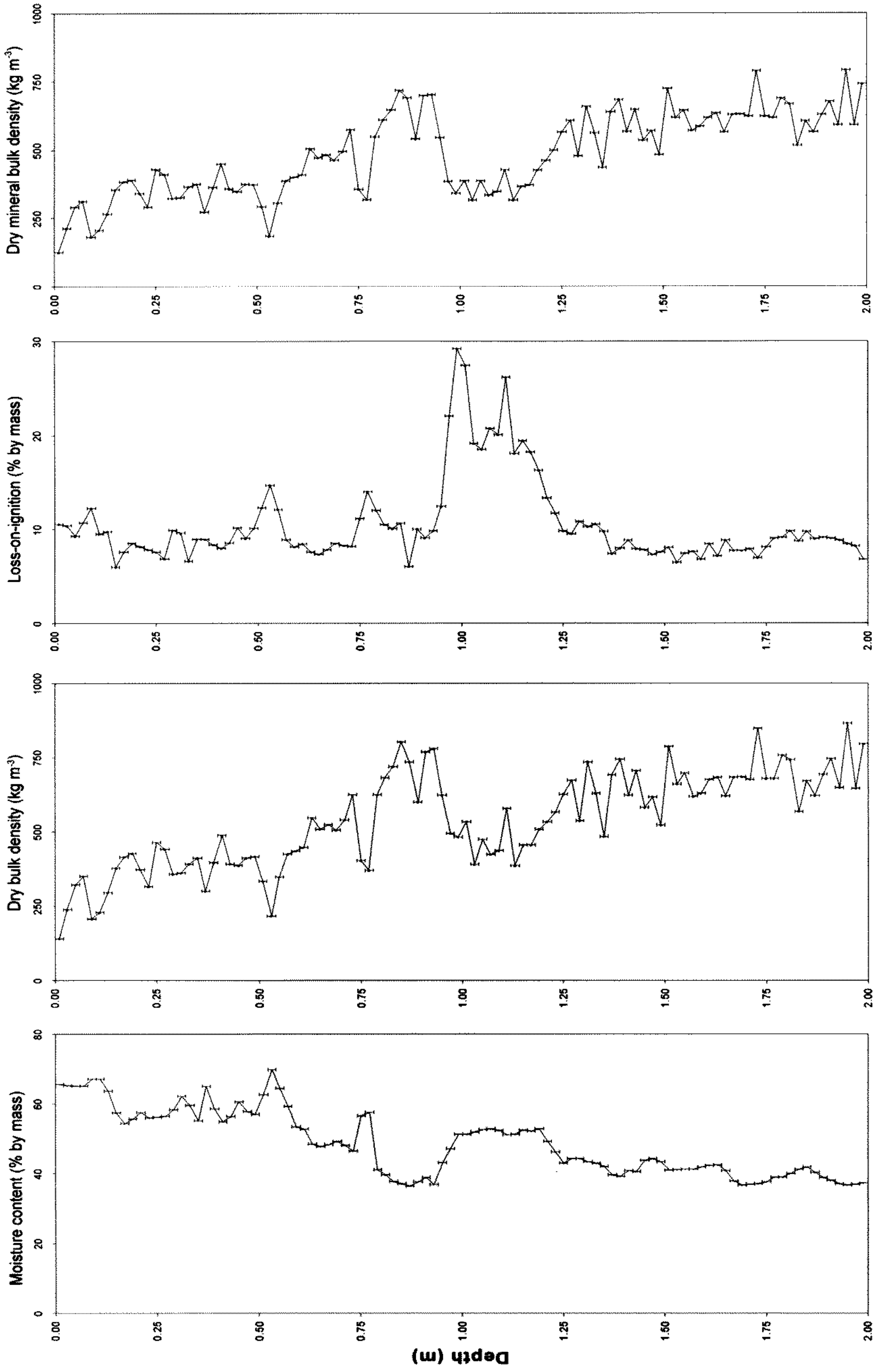
6.2.2 Moisture content

As the lake sediments were saturated at the time of sampling, measurements of moisture content in core TCA9b may be considered a surrogate of sediment porosity. Moisture content tends to decrease with depth (Figure 6.5). Superimposed on this trend are numerous well defined peaks and troughs. The decrease in porosity with depth is probably largely a consequence of compaction; the covariance of moisture content with dry bulk density supports this thesis (section 6.2.3) (Figure 6.6) (Appendix 2). From 2.00 m to 1.78 m, the moisture content alternates between a relatively constant value of *c.* 40% (2.00 m to 1.90 m, 1.62 m to 1.50 m) and short, sharp rises and decreases in water content (1.90 m to 1.62 m, 1.50 m to 1.36 m). Moisture content increases in the sediments from 39.2% at 1.36 m to 52.8% at 1.18 m, with values remaining at *c.* 52% from 1.18 m to 0.98 m.

Above 0.98 m, moisture content decreases to 37.0% at 0.92–0.94 m, with overlying values remaining below 40% until 0.80 m. From 0.80 m to the top of the core, moisture content tends to increase, with the maximum value for the core being recorded at 0.54–0.52 m (69.7%). Values fluctuate substantially through these sediments, with several sharp peaks at 0.76–0.78 m (57.5%), 0.52–0.54 m (69.7%), 0.36–0.38 m (64.9%), 0.30–0.32 m (62.1%) and 0.10–0.08 m (67.1%). A sharp rise in moisture content characterises the top 0.10 m of sediment, with values in excess of 65%.

6.2.3 Dry bulk density

Values of bulk density from core TCA9b vary from a minimum of 140.02 kg m^{-3} at 0.02–0.00 m to a maximum of 864.16 kg m^{-3} at 1.96–1.94 m (Figure 6.5) (Appendix 2). The data set exhibits substantial variability between adjacent samples, which may in part be an artefact of the constant-volume basis used to calculate the dry bulk density. Difficulties experienced in sub-sampling the sediment core, particularly with organic-rich samples, resulted in some sub-samples that were not exactly 0.02 m thick, as assumed by the dry bulk density calculation. Dry bulk density generally increases with depth through the sequence. At the base of core TCA9b, the dry bulk density was more than six times greater than at the top of the core. Between 2.00 m and 1.20 m, dry bulk density fluctuates between *c.* 500 and 860 kg m^{-3}. Values decrease over the next 0.12 m to a minimum of 385.34 kg m^{-3} at 1.14–1.12 m, and then remain between *c.* 400 and 500 kg m^{-3} until 0.96–0.94 m.


Dry mineral bulk density (kg m^{-3})
Loss-on-ignition (% by mass)
Dry bulk density (kg m^{-3})
Moisture content (% by mass)
Depth (m)

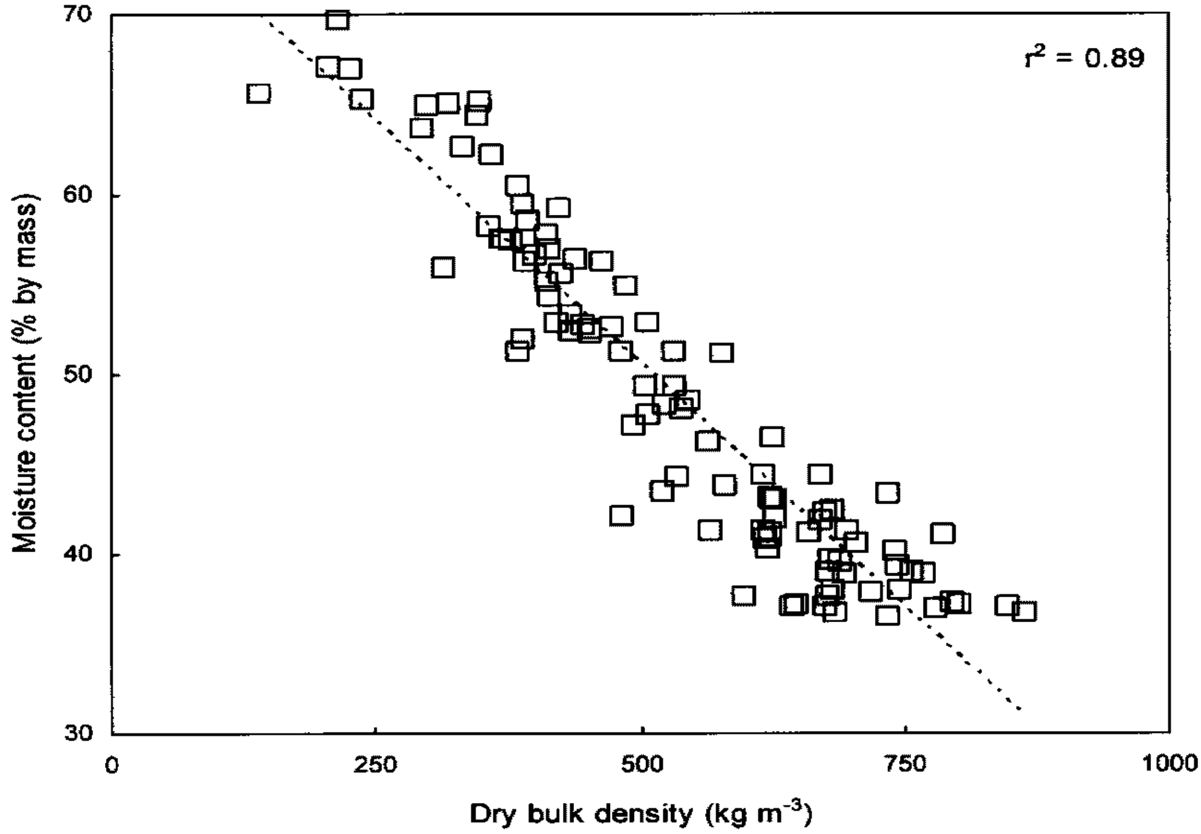


Figure 6.6 The variation of sediment moisture content as a function of dry bulk density, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

Values rise sharply above 0.96 m, remaining between 700 and 800 kg m^{-3} to 0.84 m. Above this depth, dry bulk density values generally decrease to the top of the core, although significant variation between adjacent samples was recorded. Particularly low dry bulk densities were recorded at 0.78–0.76 m (396.63 kg m^{-3}), 0.74–0.72 m (401.65 kg m^{-3}) and 0.54–0.52 m (216.54 kg m^{-3}), while individual peak values were recorded at 0.74–0.72 m (642.92 kg m^{-3}), 0.64–0.62 m (546.49 kg m^{-3}), 0.40–0.38 m (497.21 kg m^{-3}) and 0.26–0.24 m (463.84 kg m^{-3}).

6.2.4 Loss-on-ignition

The downcore variation in loss-on-ignition (LOI_{430}) from core TCA9b is presented in Figure 6.5 (see also Appendix 2). This varies from a very minor organic component of only 5.9% at 0.16–0.14 m to a maximum of 29.2% at 1.00–0.98 m. This maximum value was generated in part by the inclusion of a large piece of root wood (*Eucalyptus* genus) in the sample. Excluding the wood fragment, the corrected LOI_{430} value for this depth was approximately 22%. The underlying sample, 1.02–1.00 m, recorded a similar LOI_{430} value, 27.5%. From 2.00 m to 1.26 m, loss-on-ignition values remain approximately constant. Above 1.26 m, LOI_{430} increases to over 25% at 1.12–1.10 m depth.

After decreasing to 18.5% at 1.06 m, values rise to a maximum of nearly 30% at 1.02–0.98 m. Above this, LOI_{430} values sharply decrease to 5.96% at 0.88–0.80 m. Above 0.86 m, LOI_{430} values form a very well defined sequence of peaks and troughs with very few outlying data points. This trend continues until the top of the sediment sequence with LOI_{430} values typically remaining below 10%.

6.2.5 Dry mineral bulk density

The downcore variation in of dry mineral bulk density is presented in Figure 6.5 (see Appendix 2 for the data). There is very little difference between the pattern of dry mineral bulk density and that recorded for dry bulk density. This is confirmed by a bivariate plot of the two properties (Figure 6.7). A linear correlation of the two data sets generates an r^2 value of 0.99.

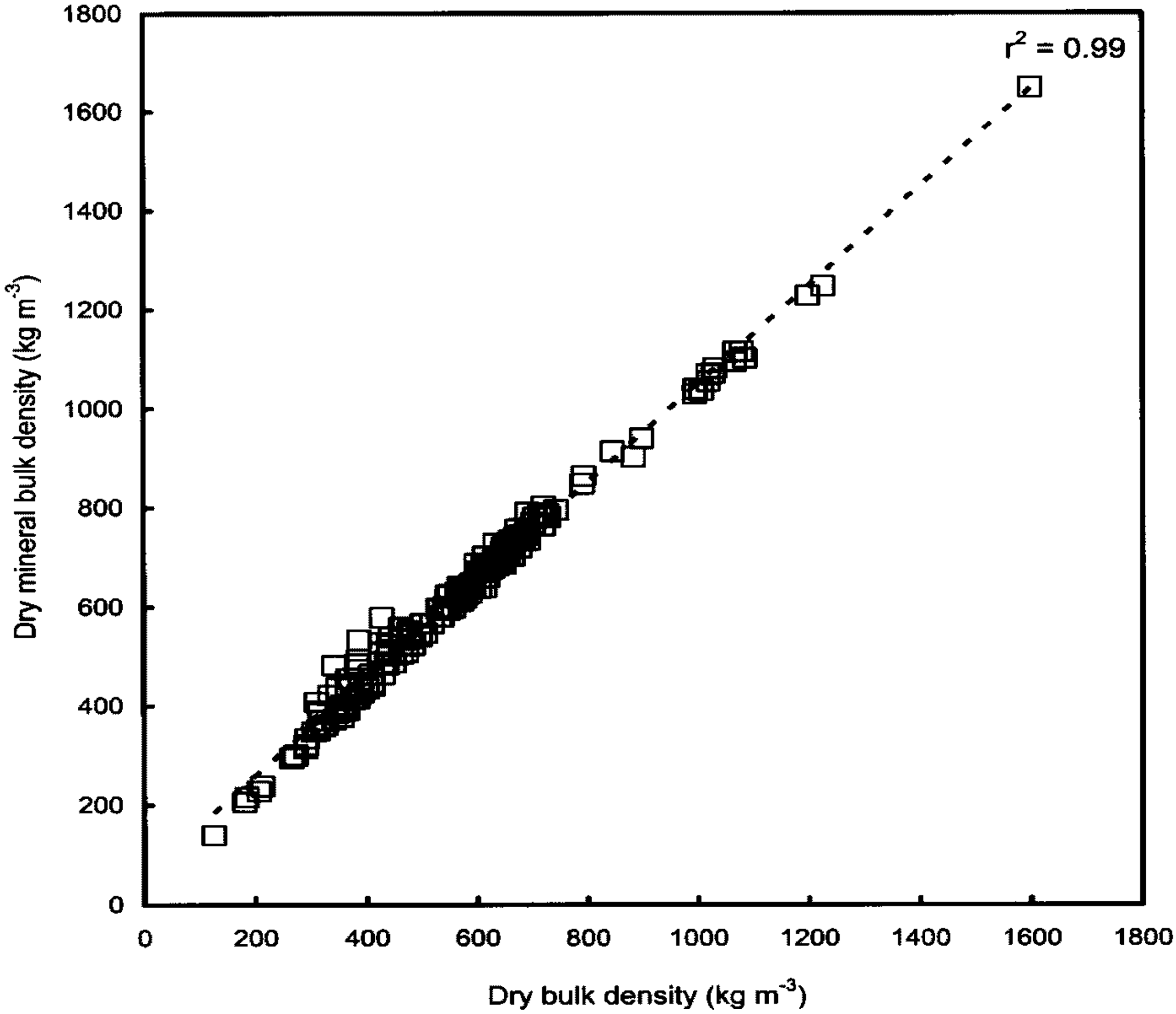


Figure 6.7 The variation of dry mineral bulk density as a function of dry bulk density, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

6.2.6 Mass-specific magnetic susceptibility

Values of mass-specific magnetic susceptibility (MSMS) for core TCA9b vary considerably with depth (Figure 6.8) (Appendix 3), with the lowest values of <10 x 10^{-8} m^3kg^{-1} occurring in the basal portion of the core and the maximum value of 379.2 x 10^{-8} m^3kg^{-1} recorded at 0.08–0.06 m. Figure 6.8 also shows the downcore MSMS values for the core on a logarithmic scale, enhancing the detail of the low susceptibility part of the record.

One of the most striking features of the profile is the long sequence of low and near–constant MSMS values in the bottom 0.70 m of the core. The mean susceptibility for this segment of the core is 7.1 x 10^{-8} m^3kg^{-1}. At 1.28–1.26 m, MSMS increases sharply by nearly 50% to 13.8 $\pm$ 0.7 x 10^{-8} m^3kg^{-1}. Above this point values remain relatively constant to a depth of 0.94–0.92 m. The susceptibility of the sediments above this depth increases sharply, doubling to *c.* 30 x 10^{-8} m^3kg^{-1}.

Susceptibility readings remain between 30 and 35 x 10^{-8} m^3kg^{-1} for the overlying *c.* 0.14 m of sediment. Between 0.64 m and the top of the core there is a series of peaks in susceptibility, increasing in magnitude towards the top of the sequence: at 0.50–0.58 m (61.5 $\pm$ 0.0 x 10^{-8} m^3kg^{-1}), 0.42–0.40 m (87.6 $\pm$ 0.8 x 10^{-8} m^3kg^{-1}), 0.18–0.16 m (163.5 $\pm$ 1.3 x 10^{-8} m^3kg^{-1}) and 0.08—0.06 m (379.2 $\pm$ 1.5 x 10^{-8} m^3kg^{-1}). The MSMS of the uppermost sample from core TCA9b (0.02–0.00 m) is 172.4 $\pm$ 0.0 x 10^{-8} m^3kg^{-1}.

6.2.7 Comparison of the magnetic susceptibility of cores TCA9a and TCA9b

The comparison of the downcore patterns of magnetic susceptibility from core TCA9a (*in situ* measurement of VSMS) and TCA9b (MSMS measurements made on sub-sampled, air-dried sediments) shows the two profiles to be quite similar in shape (Figure 6.9). Differences in the magnitude of the two signals are explained by the different measurement techniques used, with the MSMS measurements being corrected for sample density.

The susceptibility record of core TCA9b is significantly compressed when compared with that obtained from core TCA9a. This is thought to be due to the different coring procedures used to obtain the cores, with the additional force used to collect TCA9b greatly compressing the sedimentary sequence. This considered, comparing the two records, a pattern of peaks and troughs common to both signals can be recognised, allowing a convincing correlation of the two to be made (Figure 6.9).

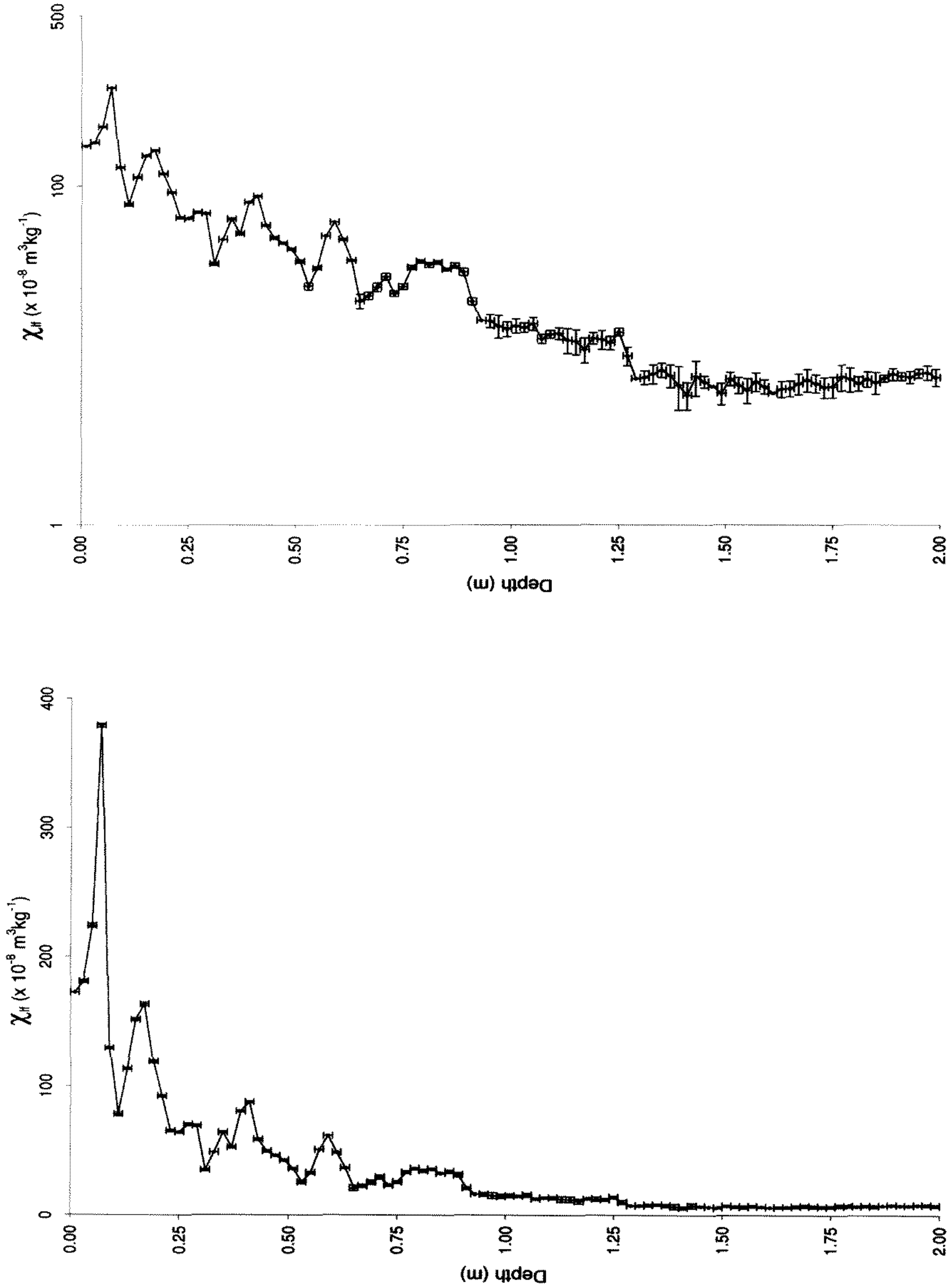


Figure 6.8 Downcore variations in mass-specific magnetic susceptibility, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 95% confidence limits for each measurement.

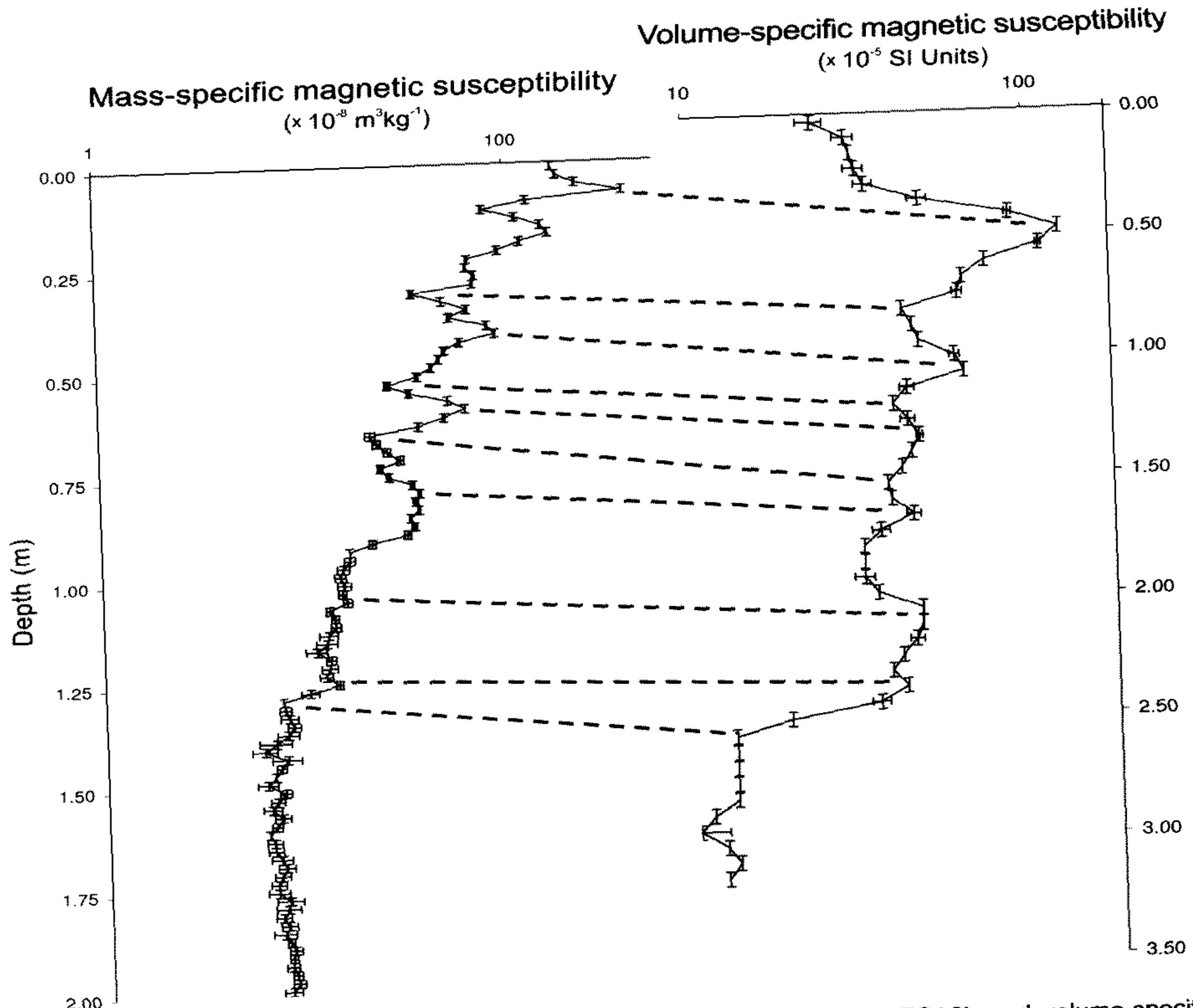


Figure 6.9 Downcore variations in mass-specific magnetic susceptibility, core TCA9b and volume-specific magnetic susceptibility, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 95% confidence limits for each measurement. Vertical axes of each chart represent depth (in metres) below the sediment surface.

6.2.8 Frequency-dependant magnetic susceptibility

The downcore variation in frequency–dependant magnetic susceptibility (χ_{fd}) for core TCA9b is presented in Figure 6.10 (Appendix 3). The data set was corrected for spurious negative values. Such values usually occur due to weakly magnetic samples generating results that are below the instrumental resolution of the Bartington MS2 meter. As Dearing (1999) has noted, the highest resolution achievable with the meter is 0.1 x 10^{-5} SI units, and therefore $K_{lf} - K_{hf}$ values $\leq$ 0.4 x 10^{-5} SI units may be regarded as lying below detection limits. Those $K_{lf} - K_{hf}$ values $\leq$ 0.4 are thus considered to be below detection limits and have been assigned zero values (that is, $[K_{lf} - K_{hf}] \leq 0.4 = 0$). The measurement errors from both the χ_{lf} and χ_{hf} data were propagated through the calculation of χ_{fd}, and are represented as horizontal error bars of two standard deviations from the mean.

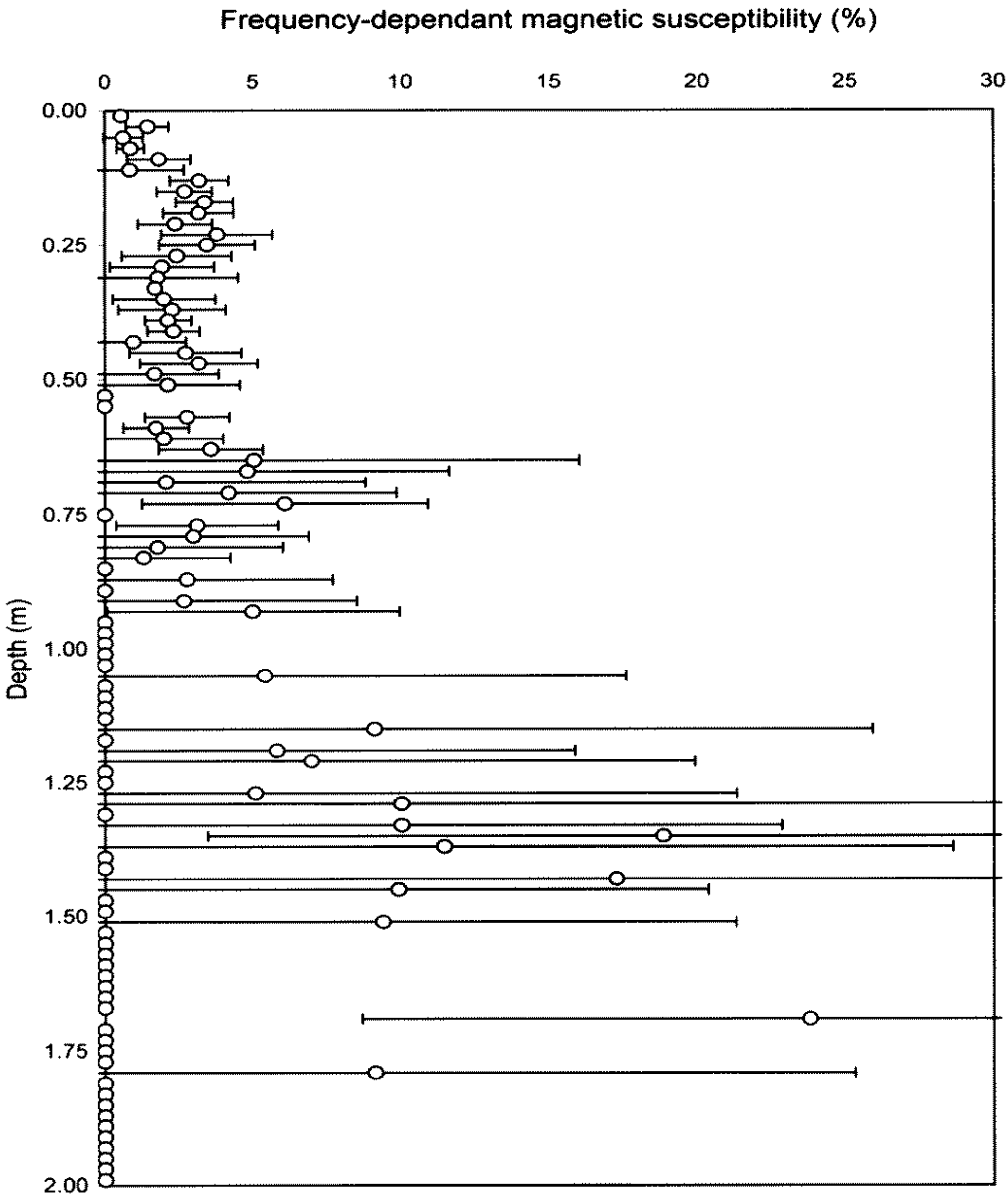


Figure 6.10 Downcore variations in frequency-dependant magnetic susceptibility, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

From 2.00 to 1.46 m, χ_{fd} values are typically equal to zero. Three measurements above zero were recorded in this section: 1.80–1.78 m (9.1%), 1.70–1.68 m (the maximum for the core, 23.8%) and 1.52–1.50 m (9.4%). Above 1.44 m, the number of non-zero values increases, with determinations alternating between zero and 10–20%. It should be noted that these high values are probably spurious given the errors involved. Values of χ_{fd} are typically zero from 1.28 to 0.94 m. Five samples generated values of between 5 and 10%, but have significantly large errors associated with them. From 0.84 m to 0.64 m, χ_{fd} values alternate between 0 and *c.* 6%, with considerably smaller errors than those recorded lower down the core. The sediments from 0.52 m to 0.12 m recorded values around the core's mean of 2.59%. These values reach a maximum of 3.79 ± 1.31% at 0.24–0.22 m. Above 0.12 m, χ_{fd} values decrease from *c.* 3% to 0.56% at the top of the core.

6.2.9 Natural remanent magnetisation

The natural remanent magnetisation (NRM) measurements from core TCA9b are presented in Figure 6.11 (Appendix 4). The downcore pattern of NRM values broadly resembles that of the MSMS measurements (Figure 6.8), with magnetically weak samples occupying the basal sediments, a sharp rise above this 'background' level and several distinctive peaks and troughs of magnetisation in the top part of the core. From 2.00 m to 1.28 m values rarely exceed 0.5 x 10^{-6} A m^2kg^{-1}, with a slight peak at 1.62–1.60 m (2.20 ± 0.14 x 10^{-6} A m^2kg^{-1}) being the exception. The first increase above this background level begins at 1.26–1.24 m (0.02 m above the similar rise in MSMS values). NRM values then steadily increase upwards to 4.03 ± 0.03 x 10^{-6} A m^2kg^{-1} at 1.12–1.10 m. Values decrease slightly above this point and, following several small fluctuations, rise sharply above 0.86–0.84 m to a peak of 32.76 ± 0.18 x 10^{-6} A m^2kg^{-1} at 0.82–0.80 m. This point is part of a plateau of elevated values identified in the downcore MSMS chart, though in that sequence the value is not so exceptionally large.

Above 0.80 m, four significant peaks occur corresponding with features at the same depths in the MSMS sequence: at 0.60–0.58 m (13.0 ± 0.07 x 10^{-6} A m^2kg^{-1}), 0.42–0.40 m (13.0 ± 0.17 x 10^{-6} A m^2kg^{-1}), 0.18–0.16 m (24.9 ± 0.82 x 10^{-6} A m^2kg^{-1}) and 0.08–0.06 m (66.16 ± 8.59 x 10^{-6} A m^2kg^{-1}). Unlike the MSMS measurements, NRM values increase markedly at the top of the core.

6.2.10 Anhysteretic remanent magnetisation

The downcore trends in anhysteretic remanent magnetisation (ARM) and mass-specific anhysteretic remanent magnetisation (χ_{ARM}) are presented in Figure 6.12 (Appendix 4). As the patterns are identical (the values differ by a constant factor), only the χ_{ARM} measurements will be discussed here.

As with MSMS and NRM, the bottom part of the core is characterised by very weak magnetic properties (in this instance, values are <0.015 x 10^{-5} m^3kg^{-1}). Mass-specific ARM values begin to increase slowly at 1.60–1.58 m, with values increasing to 2.1 ± 0.02 x 10^{-7} m^3kg^{-1} by 1.30–1.28 m. Values rise sharply above 1.26 m (paralleling the rise in MSMS measurements), gradually increasing to 7.1 ± 0.07 x 10^{-7} m^3kg^{-1} at 0.94–0.92 m. Above 0.92 m, χ_{ARM} increases rapidly, and a well-defined peak of 29 ± 0.1 x 10^{-7} m^3kg^{-1} occurs at 0.76–0.74 m. Such a peak is not found in the MSMS or NRM data; indeed the MSMS of sediments at 0.76–0.74 m is a local minimum.

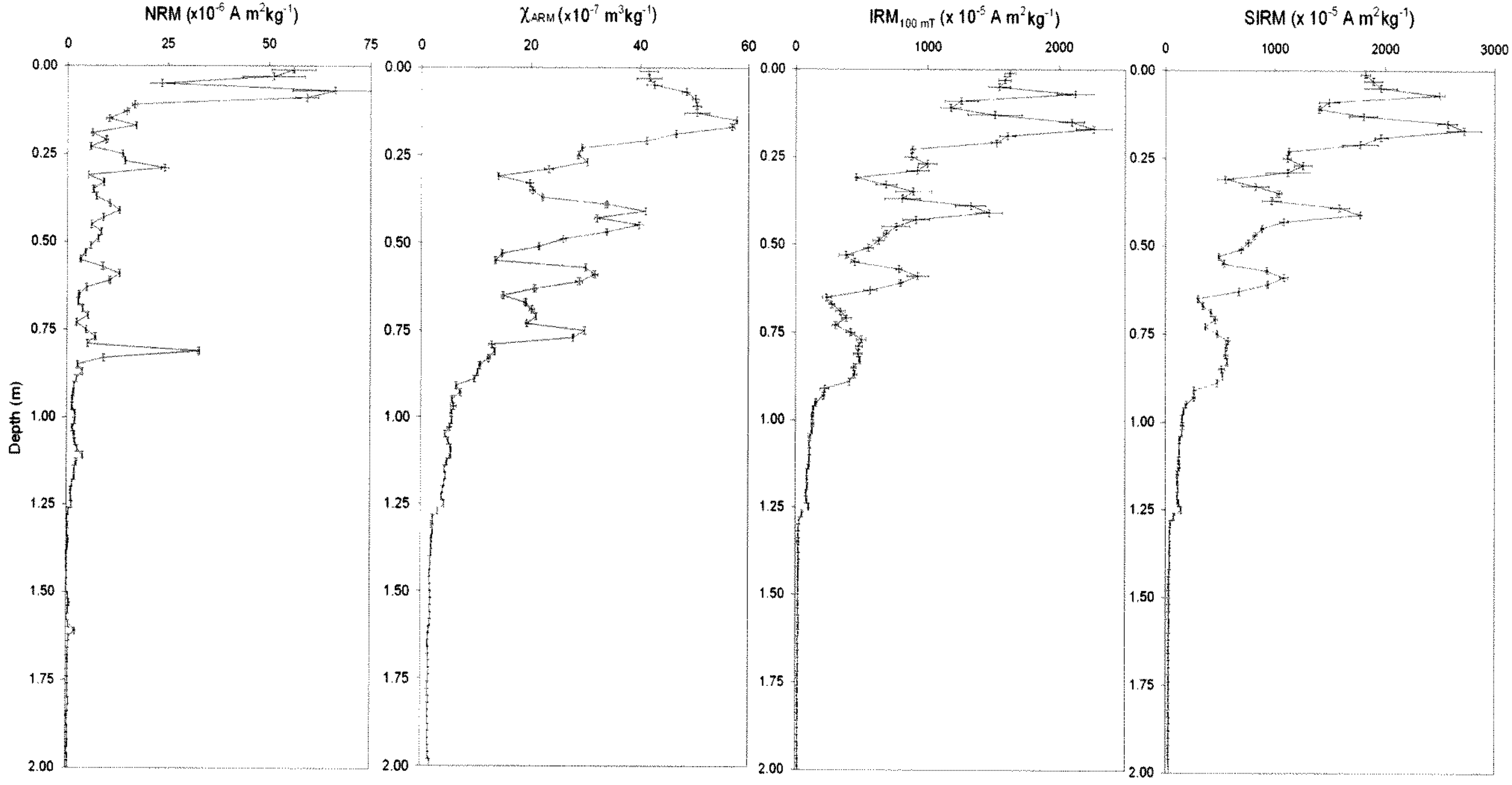


Figure 6.11 Downcore variations in NRM, χ_{ARM}, IRM$_{100\ mT}$ and SIRM, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 95% confidence limits for each measurement.

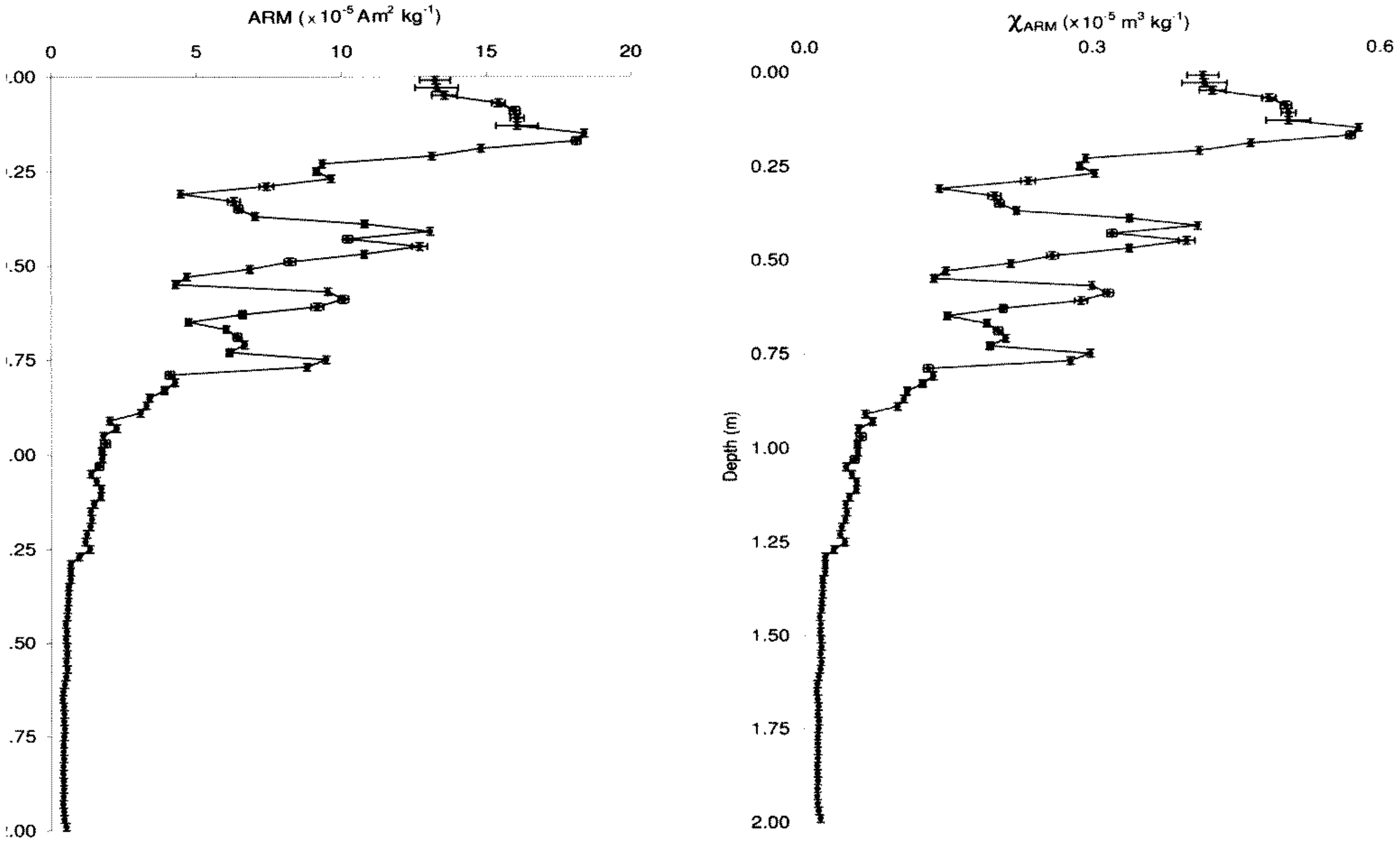


Figure 6.12 Downcore variations of ARM and χ_{ARM}, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 95% confidence limits for each measurement.

The remainder of the sequence is similar to those of MSMS and NRM, but with only three main peaks, two of which are clearly defined and common to the other data sets. Peaks at 0.60–0.58 m (31.6 ± 0.25 x 10^{-7} m^3kg^{-1}) and 0.42–0.40 m (41.0 ± 0.17 x 10^{-7} m^3kg^{-1}) are quite similar to those in the MSMS and NRM profiles. However, the 0.42–0.40 m peak is broader, with similarly high values recorded down to 0.44 m. The third main peak lies at 0.16–0.14 m (57.8 ± 0.05 x 10^{-7} m^3kg^{-1}), which is not common to the MSMS or NRM data. Above this depth, values decrease in two step-wise segments to the top of the core. No peak is found at 0.08–0.06 m, where high values are recorded in the MSMS and NRM data.

6.2.11 Isothermal remanent magnetisation

The downcore trend in isothermal remanent magnetisation ($IRM_{100\ mT}$) for core TCA9b (Figure 6.11) (Appendix 5) is nearly identical to the pattern displayed for χ_{lf}. Basal values are typically <30 x 10^{-5} A m^2kg^{-1}, while the maximum value for the core is recorded at 0.18–0.16 m (2264.9 ± 17.4 x 10^{-5} A m^2kg^{-1}).

The only significant differences in the two patterns are that the three peaks that occur at 0.60–0.58 m, 0.42–0.40 m and 0.18–0.16 m are more pronounced in the $IRM_{100\ mT}$ data set. The peak at 0.18–10.16 m, the maximum $IRM_{100\ mT}$ value for the core, is not as pronounced in the χ_{lf} data. The uppermost peak at 0.08–0.06 m is significantly contracted compared to that recorded in the χ_{lf} data.

6.2.12 Saturated isothermal remanent magnetisation

The saturated isothermal remanent magnetisation (SIRM) measurements for core TCA9b are presented in Figure 6.11 (Appendix 5). The pattern of downcore SIRM values is near identical to that recorded by the IRM_{100mT} data. The most significant differences occur at the synchronous peaks, where the SIRM values are relatively much higher than the IRM_{100mT} values.

6.2.13 χ_{ARM}/χ_{lf}

Values of χ_{ARM}/χ_{lf} are quite low at the base of core TCA9b, and increase slowly until 0.80–0.78 m (Figure 6.13) (Appendix 6). Within this basal unit, χ_{ARM}/χ_{lf} values fluctuate occasionally, with peaks at 1.74–1.72 m, 1.42–1.40 m, 1.18–1.16 m and 0.94–0.92, and minima at 1.70–1.64 m, 1.36–1.34 m, 1.06–1.04 and 0.90–0.80 m. Above 0.78 m, the ratio shifts dramatically, with χ_{ARM} increasing relative to χ_{lf}. By 0.76–0.74 m, the ratio has increased by over three times to 11.6. Above this depth, values of χ_{ARM}/χ_{lf} decrease in a step-wise fashion, punctuated by peaks at 0.68–0.66 m, 0.46–0.44 m and 0.12–0.10 m. Above 0.10 m, the χ_{ARM}/χ_{lf} ratios are similar to those of the lower 1.2 m of the core.

6.2.14 S-ratio ($IRM_{100\ mT}$/SIRM)

The TCA9b S-ratio record consists of three well-defined components; the lower *c.* 0.75 m of relatively low, noisy values, the top *c.* 0.90 m of higher values, and a transition between the two (*c.* 1.25 m to 1.10 m) (Figure 6.13) (Appendix 6). The S-ratio fluctuates repeatedly between 0.5 and 0.7 from 2.00 m to *c.* 1.50 m depth, with the lowest value for the core (0.55) recorded at 2.00–1.98 m. Above 1.48 m, values stabilise and steadily decrease to 0.57 at 1.36–1.34 m. Values rise significantly above 1.36 m, exceeding the mean S-ratio for the bottom 0.75 m of the core (0.65) at 1.28–1.26 m. From 1.10 m, S–ratios fluctuate between *c.* 0.80 and *c.* 0.90 with a series of peaks and troughs. Local maxima were recorded at 0.90–0.88 m, 0.76–0.74 m, 0.60–0.58 m, 0.46–0.44 m, 0.36–0.38 m, 0.22–0.20 m, 0.08–0.06 m and 0.02–0.00 m (the core's maximum S-ratio, 0.89).

6.2.15 High-field IRM (SIRM–IRM$_{100mT}$)

The high-field or 'hard' component of the IRM (HIRM)(Maher *et al.*, 1999; Crockford and Willet, 2001) from core TCA9b is presented in Figure 6.13 (Appendix 6). This measure represents the magnetically coarser component of the bulk sediment. This will only become magnetised in the maximum (SIRM) field imparted. The downcore trend is nearly identical to that exhibited by the IRM and SIRM data (sections 6.2.6 and 6.2.7), though two main differences were recorded. First, the initial significant rise above the long-term, low susceptibility background of the bottom *c.* 0.75 m is more pronounced in the 'hard' IRM plot, suggesting a significant contribution of magnetically 'harder' material at this depth. Above this first peak (1.26–1.24 m), 'hard' IRM values decrease for the next *c.* 0.15 m of overlying sediment. The second difference in the plot is that the peak features recorded are more exaggerated than those from either the IRM or SIRM data. This again suggests that at these points of peak susceptibility there is probably a greater contribution of magnetically 'hard' material in the sediment.

6.2.16 SIRM/χ_{lf}

The SIRM/χ_{lf} values for TCA9b increase gradually and inconsistently from the base of the core (SIRM/$\chi_{lf} \approx 0.4$ A m^{-1}) to 1.30–1.28 m (SIRM/$\chi_{lf} = 6.1$ A m^{-1}) (Figure 6.13) (Appendix 6). Values rise rapidly above this reaching a local maximum of 1.03 at 1.26–1.24 m. The 0.10 m of sediment above here records values of SIRM/χ_{lf} that fluctuate between *c.* 0.8 and 1.05 A m^{-1}. Another sharp rise occurs from 0.98–0.92 m. Values continue to increase slowly through a number of small, but well defined peaks and troughs to the core's maximum of 2.03 A m^{-1} at 0.42–0.40 m. From 0.40 m to 0.10 m three major peaks in the ratio occur at 0.28–0.26 m, 0.22–0.20 m and 0.12–0.10 m. Above here the SIRM/χ_{lf} ratio falls steeply to 0.64 A m^{-1} at 0.08–0.06 m, increasing to 1.05 at the top of the core.

6.2.17 χ_{ARM}/SIRM

The χ_{ARM}/SIRM values recorded from the base of the core to 1.64 m remain relatively stable, and consistently high at between 40 and 50 x 10^{-5} m A^{-1} (Figure 6.13, Appendix 6). Above 1.64 m, values increase steadily to 52 x 10^{-5} m A^{-1} at 1.52–1.50 m, before reaching the maximum value for the basal segment of the core at 1.48–1.46 m. A decline in χ_{ARM}/SIRM occurs above this depth, with values increasing to a localised peak at 1.34–1.32 m of similar magnitude to that recorded 0.14 m below. χ_{ARM}/SIRM values decrease rapidly above this depth, reaching a local minimum at 1.26–126 m.

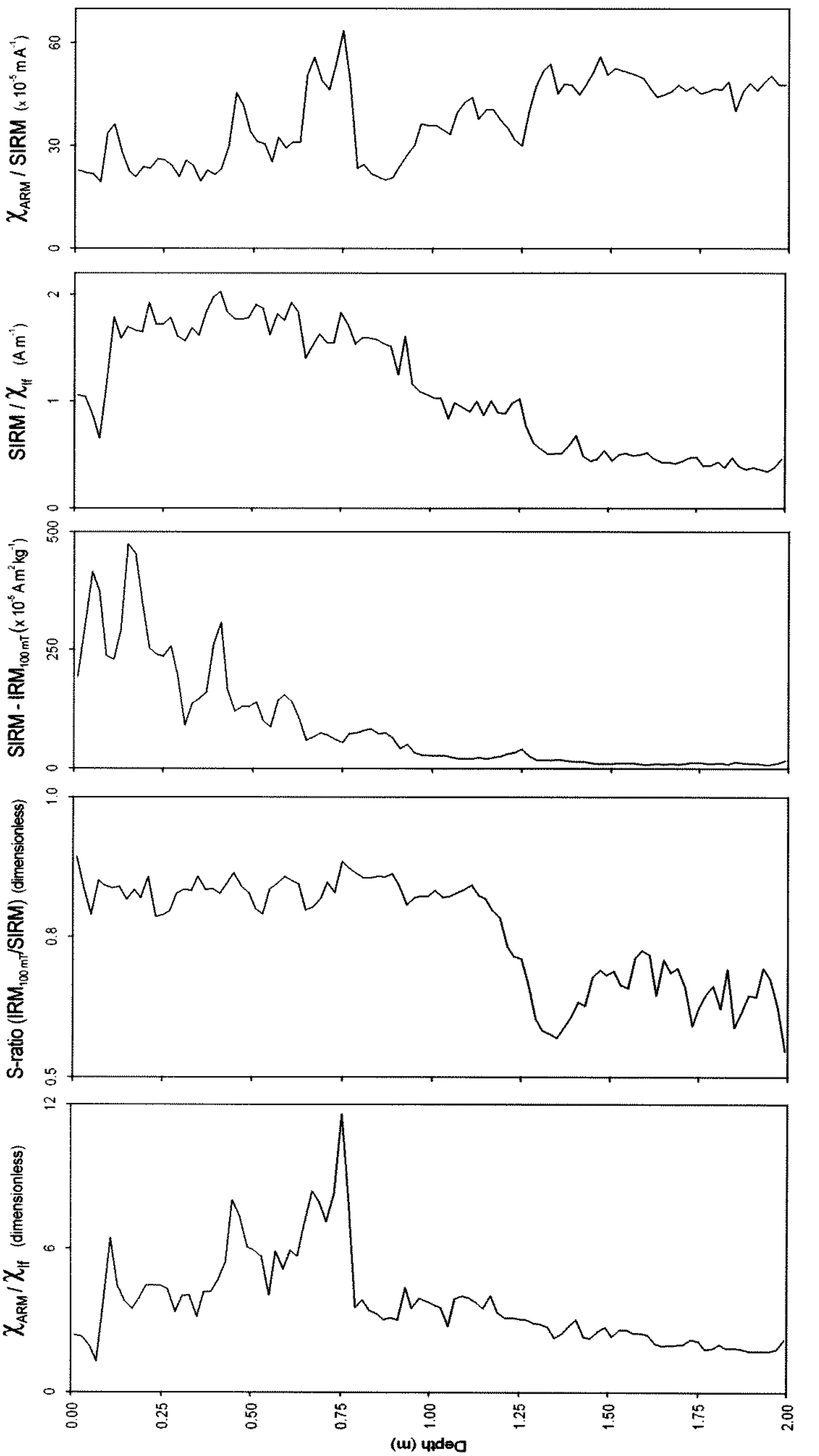


Figure 6.13 Downcore variations in χ_{ARM}/χ_{lf}, the S-ratio, SIRM-$IRM_{100 mT}$ (HIRM), SIRM/χ_{lf} and χ_{ARM}/SIRM, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

Three peaks in the ratio at 1.18–1.14 m, 1.12–1.10 m and 1.02–0.96 m are superimposed upon a longer term decrease in the χ_{ARM}/SIRM ratio that culminates at 0.88–0.86 m (20 x 10^{-5} m A^{-1}). Above 0.80 m, values of the ratio more than double to the core's maximum of 63 x 10^{-5} m A^{-1} at 0.76–0.74 m. The *c.* 0.40 m of sediment overlying this peak record a generally decreasing trend in χ_{ARM}/SIRM values; however two distinct peaks occur at 0.68–0.66 m and 0.46–0.44 m. Above each peak the ratios decrease sharply, finally stabilising at 20–25 x 10^{-5} m A^{-1} between 0.42 m and 0.16 m. A final peak in χ_{ARM}/SIRM is centred on sediments at 0.12–0.10 m, while the uppermost 0.10 m of the core records values in the low 20s (x 10^{-5} m A^{-1}). All the major peaks in χ_{ARM}/SIRM coincide with those detected in the χ_{ARM}/χ_{lf} plot suggesting that these features are controlled by χ_{ARM} rather than by any other magnetic property.

6.2.18 S-ratio versus χ_{lf}

Plotting the S-ratio for core TCA9b against the corresponding χ_{lf} data (Figure 6.14) further teases out the relationship exhibited in Figure 6.13 (S-ratio) between those sediments below and above 1.28 m in the core. Those sediments below 1.26 m all have χ_{lf} values below 10 x 10^{-8} m^3kg^{-1} and S-ratios below 0.73, and cluster linearly in the bottom left-hand corner of the chart. The three samples from 1.26 m to 1.20 m plot along a path between the basal sediment cluster and the upper cluster of points from the remainder of the core.

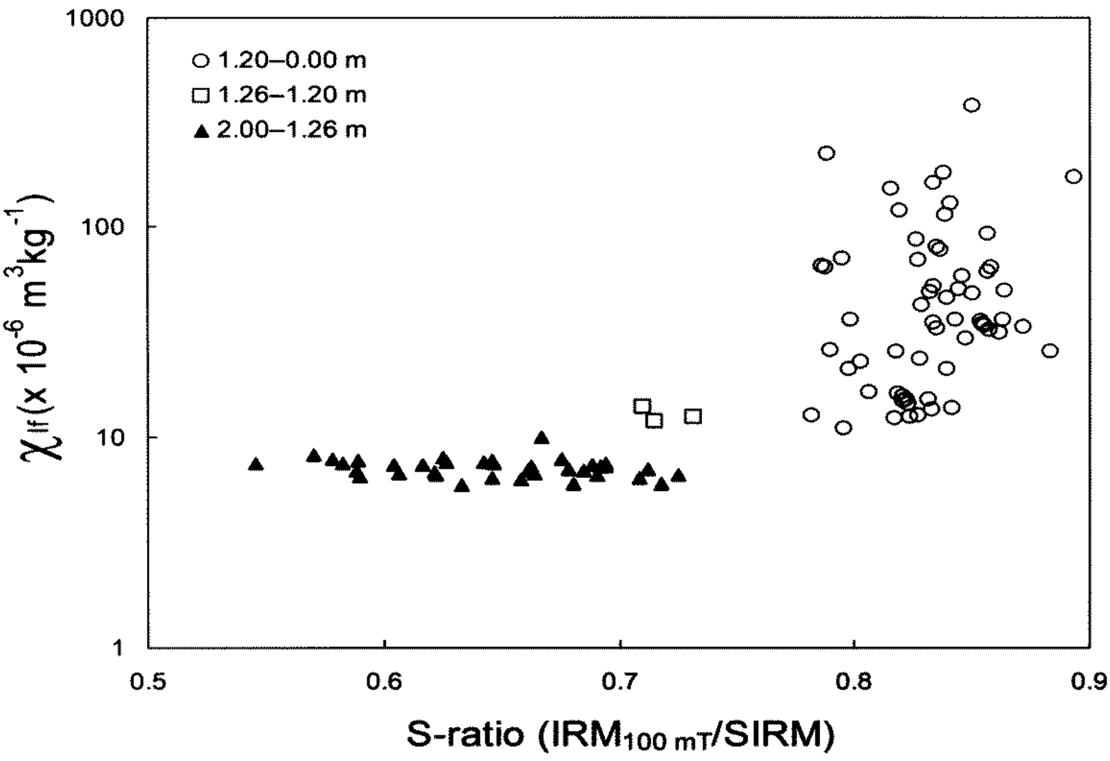


Figure 6.14 The variation of χ_{lf} as a function of S-ratio, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

6.3 Lake sediment palaeomagnetism

The downcore measurements of magnetic inclination made on cores TCA9c and TCA9d from Tocal Homestead Lagoon (section 5.3.2) are presented in Figure 6.15 (see Appendix 7 for raw NRM data). An identifiable inclination signal is broadly common to both cores, with lower inclinations towards the base of the cores, which generally steepen up-core. Repeat measurements of core TCA9c showed the sediments to have preserved a reproducible inclination record. Two inclination minima, at 3.41–3.45 m and 3.10–3.05 m, characterise the basal sediments of core TCA9c. Between 3.00 and 2.50 m, inclination values steepen sharply culminating in the steepest inclinations for the entire sequence of -80 to -85°. Above this maximum, inclination sharply becomes shallower, with the minimum value for TCA9c recorded at 1.80 m. Inclinations steadily steepen towards the top of the core, reaching -76° at the top. A broadly similar, but more complicated pattern is exhibited in the first inclination log of the replicate core TCA9d.

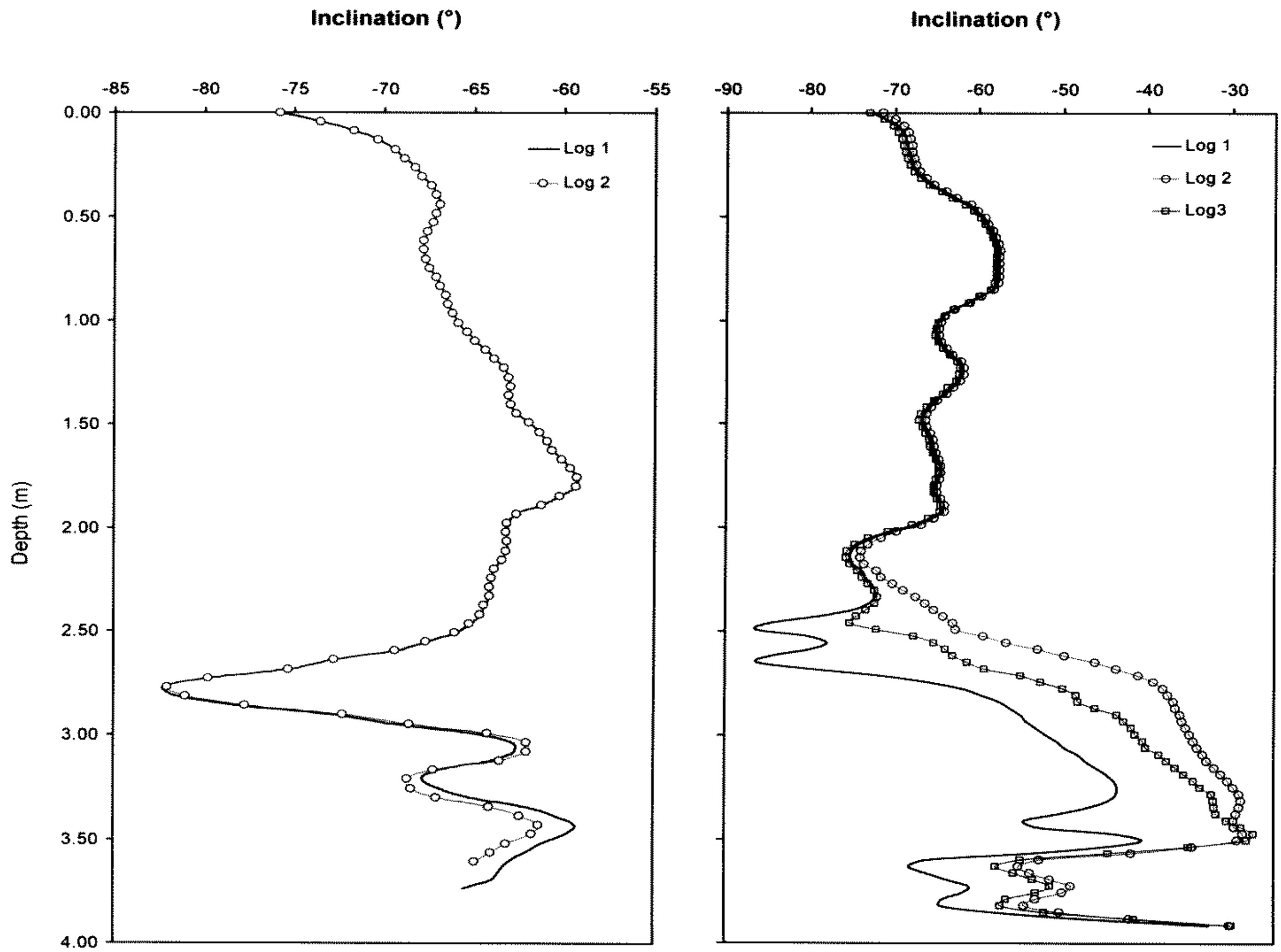


Figure 6.15 Downcore variations in magnetic inclination (the vertical component of NRM), cores TCA9c (left) and TCA9d (right), Tocal Homestead Lagoon, Paterson, eastern Australia.

Below *c.* 2.20 m, however, the pattern differs significantly between measurement runs, with inclination decreasing until 3.50 m in the second and third logs, before the three logs behave with some similarity at the base of the core. Although core TCA9d preserves a longer and more detailed inclination signal from the lake, these repeat measurements show the lower half of the sequence to be incapable of retaining a stable NRM. This may be related to the magnetic susceptibility of the basal sediments, which decreases sharply below *c.* 2.20 m, suggesting that a much lower concentration of remanence carrying minerals is present in comparison to those sediments deposited above (Figure 6.16). As both cores were taken from the same location, this curious difference in the stability of NRM signals in the two cores might be better explained by some disturbance of the sediments after coring, perhaps during extraction or during transport, causing a loss of remanent magnetisation in the lower sediments of core TCA9d.

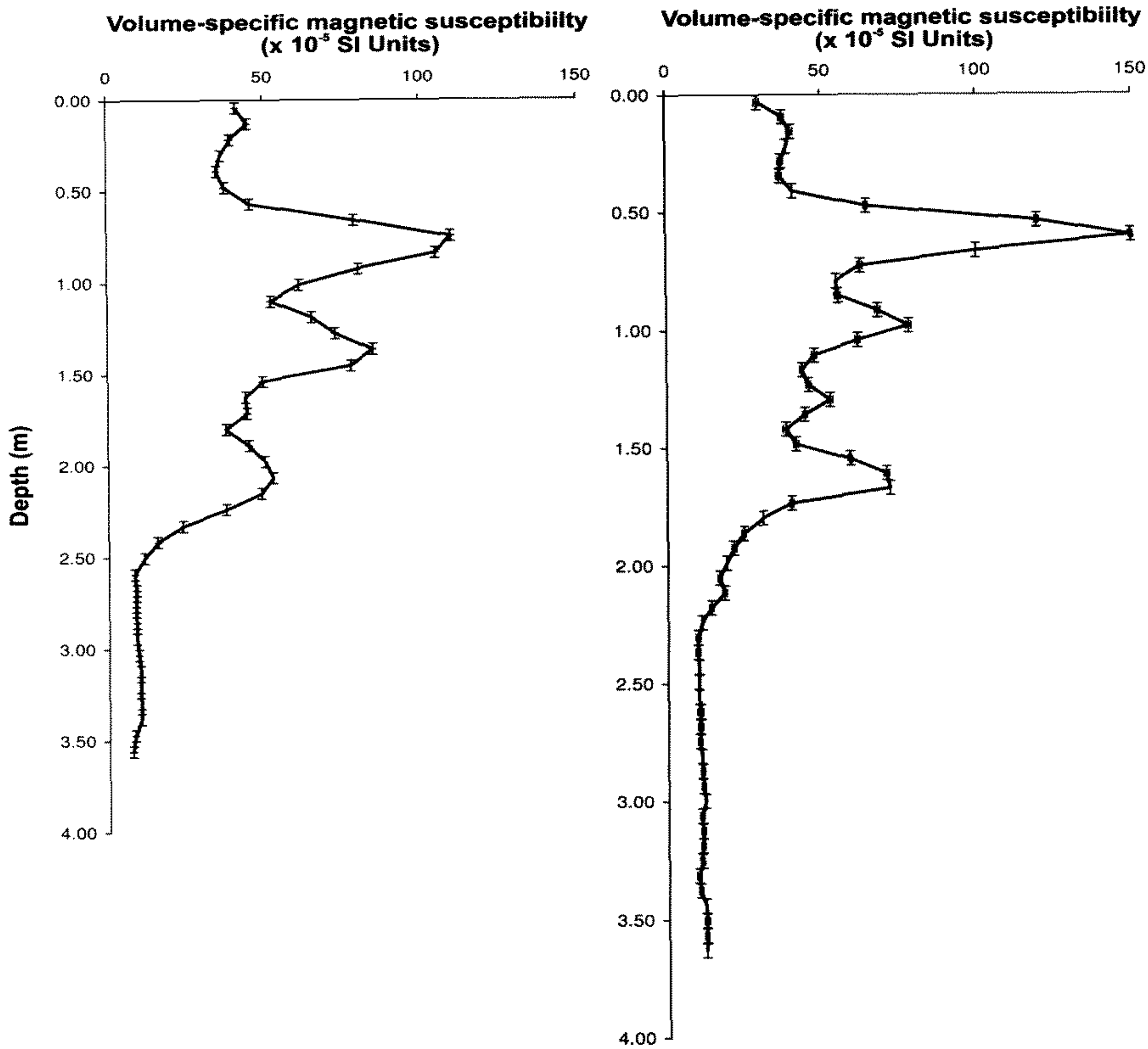


Figure 6.16 Downcore variation of volume-specific magnetic susceptibility for cores TCA9c (left) and TCA9d (right), Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sampling intervals, while horizontal error bars represent 95% confidence limits for each measurement.

6.4 Lake sediment geochemistry

6.4.1 Major element geochemistry

6.4.1.1 Introduction

The major element geochemistry of core TCA9b sediments is presented in Figure 6.17 (Appendix 8). All data are expressed as variations in elemental oxide concentrations (%) with depth (m). Each value has been corrected for the mass of sediment sampled and for the mass of material (mainly organic material) lost on combustion at 1100°C (section 5.6.1) as part of the analytical procedure. Of the 25 major elements analysed, 16 (Al, Ba, Ca, Fe, K, Mg, Mn, Na, P, S, Si, Sr, Ti, V, Zn, Zr) recorded concentrations higher than the detection limits of the procedure. Further discussion will therefore be restricted to these elements. To avoid the problems of closed data sets, which may result in variations of components of the chemical assemblage as a function of changes in the incidence of other components, the chemical concentration data were re-expressed as ratios of SiO_2, the dominant oxide in the assemblage. Six of the oxides vary little in concentration downcore (Al, Si, Ti, V, Zn, Zr) (Figure 6.17). However, Zn levels increase above *c.* 0.60 m. Another group of oxides displays a pattern of relatively low concentrations in the bottom *c.* 0.75 m of the core, with values increasing above this depth (Fe, Mg, P, Ca, Mn, S). This is similar to the pattern recorded in many of the magnetic measurements (section 6.2). Finally, the oxides of K, Ba and Sr increase upwards from 2.00 m, reaching a peak above *c.* 1.26 m, above which a notable trough feature is recorded until *c.* 1.00 m. Concentrations then increase through a series of broad peaks and troughs reaching their maximum values at or near the top of the core.

6.4.1.2 The relationship between iron and magnetic susceptibility

A bivariate plot of the magnetic susceptibility of the core TCA9b sediments versus their normalised iron (Fe_2O_3/SiO_2) content is presented in Figure 6.18. The magnetic susceptibility of these sediments increases with increasing iron content, suggesting that the magnetic signal is controlled by the concentration of iron minerals. Those sediments below 1.26 m exhibit little variation and are characterised by low iron contents and susceptibility. Those sediments above this 1.26 m display much greater iron contents and higher susceptibility, but the relationship between the two exhibits much greater variability.

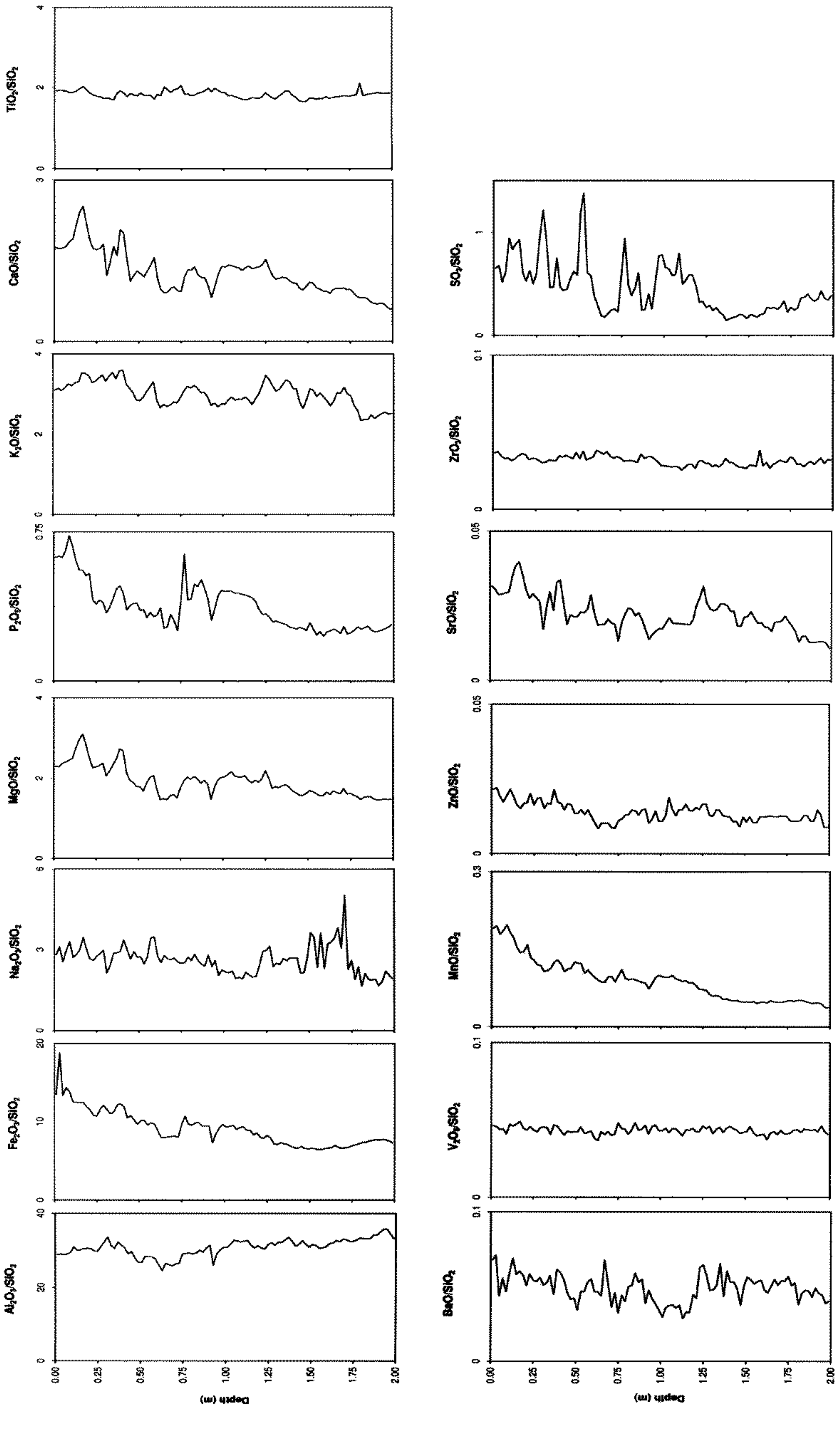


Figure 6.17 Downcore variation in the oxides of Al, Ba, Ca, Fe, K, Mg, Mn, Na, P, S, Sr, Ti, V, Zn and Zr expressed as ratios of SiO_2, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Ratios have been exaggerated by 100 times for presentation purposes.

6.4.1.3 Data analysis

Principal component analysis (PCA) utilising a covariance matrix was undertaken on the oxides of Al, Ba, Ca, Fe, K, Mg, Mn, Na, P, S, Sr, Ti, V, Zn and Zr expressed as ratios with SiO_2 for each sample from the top 2.00 m of core TCA9b (n = 100). The first two principal components explain 95.6% of the total variance in the data set. A plot of the first principal component scores versus the second principal component scores clearly discriminates between those samples from below 1.28 m in the sequence and those above (Figure 6.19). Sediments below 1.28 m display little chemical variance, implying a relatively consistent depositional environment. The linear relationship between PC1 (dominated by Ca, Mg, Fe, Mn) and PC2 (dominated by Na, Ba, S and K) in the below-1.28 m sediments suggests that increases in major elements occur uniformly, and that little individual variation in elements occurs in these sediments. Above 1.28 m the bulk sediment geochemistry changes dramatically, exhibiting much greater variability.

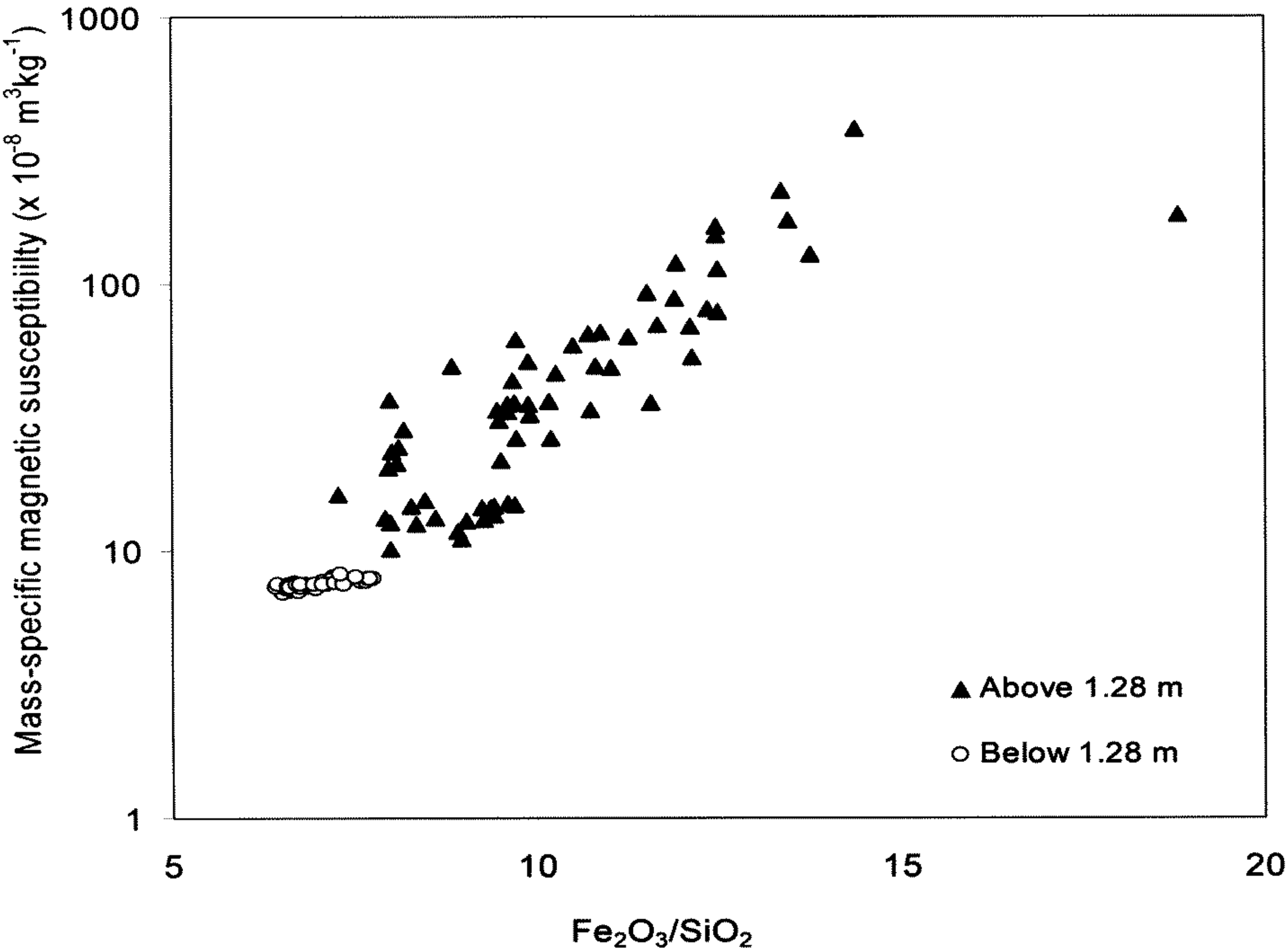


Figure 6.18 The variation of mass-specific magnetic susceptibility as a function of Fe_2O_3/SiO_2, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

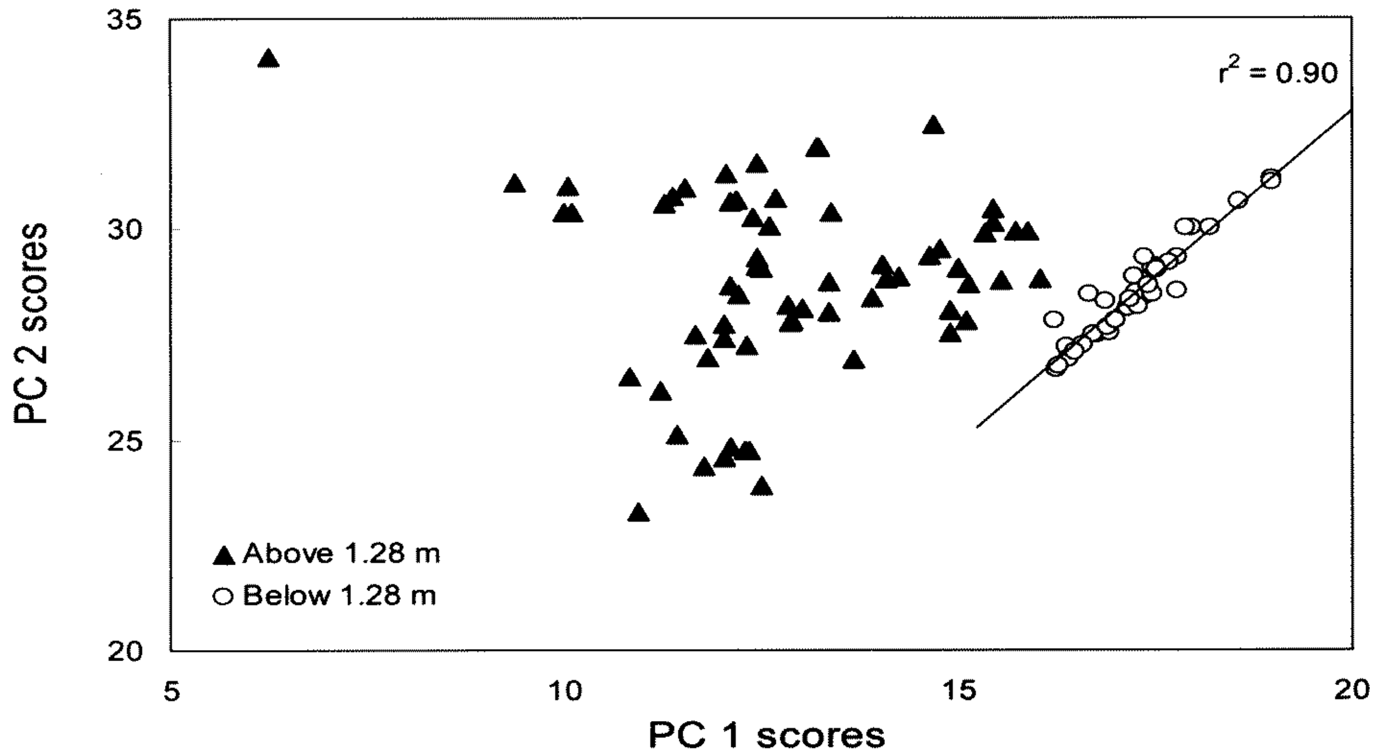


Figure 6.19 Principal component analysis (by covariance) of the oxides of 15 elements for each sample from the top 2.00 m of core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Each oxide is expressed as a ratio of SiO_2. The first two principal components explain 95.6% of the total variance in the data set.

6.4.1.4 Phosphorus

To avoid the problem of closed data sets (Aitchison, 1986), P_2O_5 concentrations have been expressed as ratios of SiO_2, Al_2O_3 and Fe_2O_3 (Figure 6.20) (Appendix 9). These three oxides make up the bulk of the sediment chemistry. The two plots P_2O_5/SiO_2 and P_2O_5/Al_2O_3 show near-identical patterns, whilst the P_2O_5/Fe_2O_3 plot shows the same, though somewhat exaggerated, pattern of peaks and troughs. The correspondence in the pattern displayed by the three ratios strongly suggests that the increases represent real changes in the concentration of phosphorus in the deposits rather than artefacts of variations in other components of the geochemical assemblages. As the major source of phosphorus in undisturbed systems is mineral input as P-Fe rich clays, detrital P should exhibit a fixed ratio with Fe (dependent on the major mineral input into a system). Changes in human input of phosphorus into the system should therefore be recorded as increases in P independent of changes in Fe.

The upcore variation of P_2O_5/Fe_2O_3 in the core shows phosphorus concentrations beginning to increase significantly with respect to Fe above 1.74 m (Figure 6.20). A further increase in phosphorus independent of Fe begins at 1.28–1.26 m, mirroring similar rises recorded in the P_2O_5/SiO_2 and P_2O_5/Al_2O_3 ratios. In addition, the P_2O_5/Fe_2O_3 data alone show a sharp rise in P in the sediments independent of Fe 1.26–1.24 m. The greatest value of P_2O_5/Fe_2O_3 is recorded at 0.78–0.76 m, above which values decline rapidly. All three ratios show reduced phosphorus levels above 0.74 m, similar to those documented from 1.26 to *c.* 0.90 m. Sediments overlying this depth show increasing phosphorus concentrations. Maximum values for the core occur at 0.10–0.08 m (0.12–0.08 m in the P_2O_5/Fe_2O_3 record). The phosphorus ratios determined by ICP–AES show near identical patterns to those shown in the XRF data (Figures 6.20 and 6.21) (Appendix 9), suggesting that the two chemical data sets are comparable. Therefore, only the ICP-MS phosphorus results will be considered for interpretation. Phosphorus levels begin to rise above about 1.50 m depth, a change that is confirmed by the P/Al and P/Fe plots. The dramatic rate of phosphorus increase from 1.50–1.48 to 1.18–1.16 m is much greater in the phosphorus versus depth plot than in the P/Al or P/Fe charts. This would suggest that the P increase in these sediments is above that associated with detrital flux.

The major peaks in phosphorus concentration correspond to similar peaks in P/Al and P/Fe (at 0.78–0.76 m, for example), showing that peaks in the concentration of phosphorus occur independently of the stable fractions (Si, Al, Fe) of the sediment chemistry. This confirms the picture provided by the corresponding XRF data. This suggests that these peaks in phosphorus record changes in phosphorus input into the lake system.

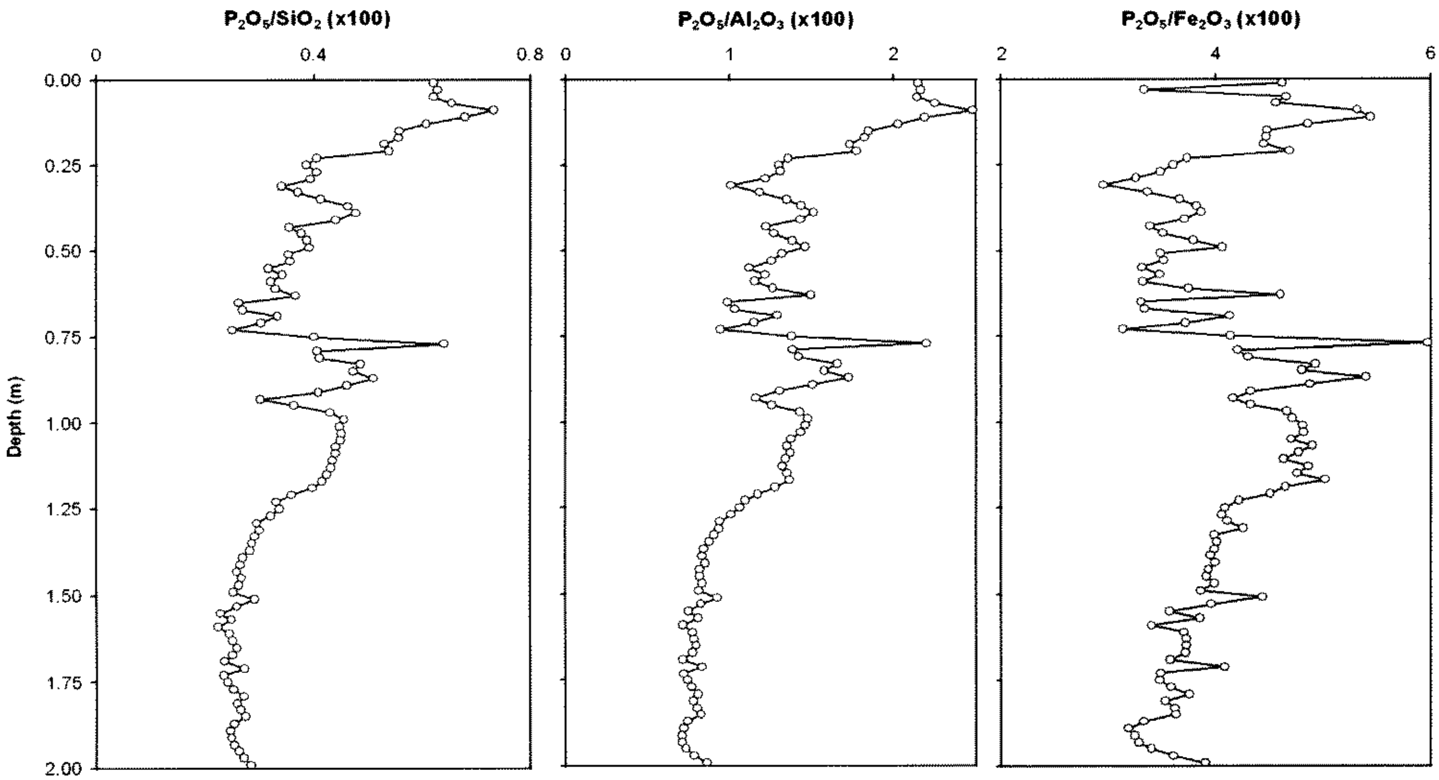


Figure 6.20 Downcore variation in P_2O_5 (expressed as ratios of SiO_2, Al_2O_3 and Fe_2O_3), core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

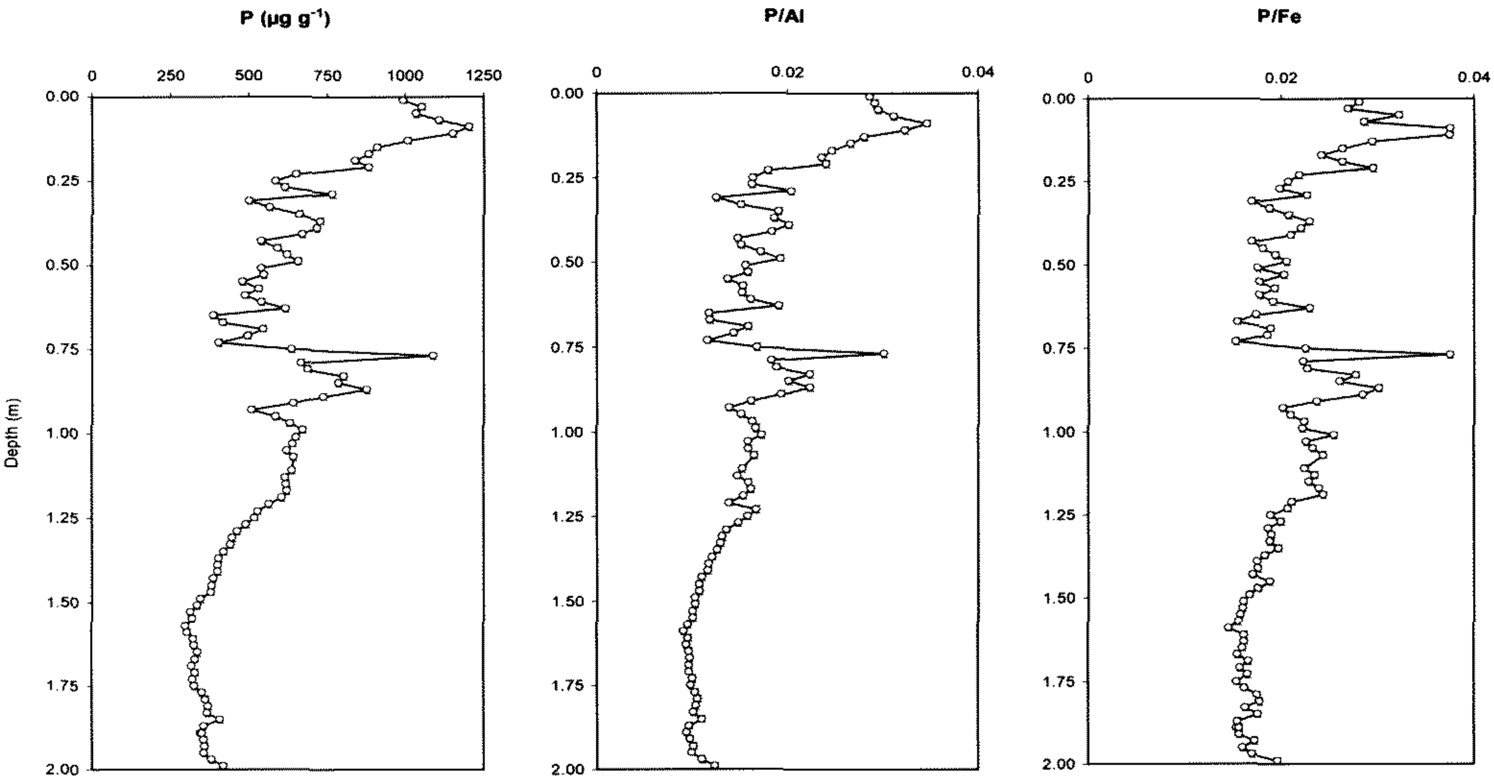


Figure 6.21 Downcore variation of P, P/Al and P/Fe, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

6.4.2 Minor element geochemistry

6.4.2.1 Introduction

The downcore variations in the concentration of Cd, Cr, Cu and Pb are presented in Figure 6.22 (see Appendix 10 for the original data).

6.4.2.2 Cadmium

The concentration of Cd in the core TCA9b sediments is very low, typically less than 1 µg g^{-1} (Figure 6.22) The overall pattern is noisy, with many high-frequency fluctuations in values. One explanation for this is that these values are close to the limits of detection of ICP-AES. The calculated 95% confidence limits on these measurements confirm that this may be likely. The concentration of cadmium in the lower 0.50 m of sediments is very low, between 0.2 and 0.4 µg g^{-1}. A prominent exception is a large spike at 1.94–1.92 m (1.1 µg g^{-1}, the maximum value for the core) and, to a lesser extent, 1.92–1.90 m (0.9 µg g^{-1}). After an anomalous zero value of 1.44–1.42 m, cadmium content slowly rises, albeit unevenly, until the top of the core. Through this steady rise in cadmium values peaks were recorded at the following depths: 1.30–1.28 m (0.8 µg g^{-1}), 0.96–0.94 m (0.8 µg g^{-1}), 0.84–0.82 m (0.7 µg g^{-1}), 0.76–0.74 m (0.7 µg g^{-1}) and 0.48–0.46 m (1.0 µg g^{-1}). Only one of these peaks consists of more than one data point (0.76–0.74 m).

6.4.2.3 *Chromium*

The concentration of Cr in core TCA9b varies little through the sequence. The bottom 0.75 m of the core is characterised by a series of broad peaks. Local maxima were recorded at 1.90–1.80 m, 1.66–1.60 m and 1.48–1.26 m. From 1.26 m, values decrease consistently to 1.14 m. Variability in Cr increases above this depth, and values fluctuate between 35 and 40 $\mu g\ g^{-1}$ for the remainder of the core. A series of minor troughs in Cr is recorded in the upper metre of sediment: from 0.86 m to 0.76 m, 0.60 m to 0.56 m, 0.26 m to 0.18 m and 0.14 to 0.08 m

6.4.2.4 *Copper*

The pattern of Cu concentration in core TCA9b is similar to that of Cr for the bottom *c.* 0.75 m of sediment. Above this depth, the behaviour of copper becomes desynchronised from that of chromium. The maximum concentration occurs at 1.26–1.24 m (31.6 $\mu g\ g^{-1}$), with values decreasing above this to the core's minimum Cu content of 13.6 $\mu g\ g^{-1}$ at 0.94–0.92 m. Sediments deposited above this depth have Cu concentrations that fluctuate through a series of narrow and spiky peaks and troughs, with concentrations remaining between 14 and 24 $\mu g\ g^{-1}$ for the remainder of the core. Peaks values in the top metre of sediment occur at the following depths: 0.80–0.78 m, 0.52–0.50 m, 0.42–0.34 m and 0.18–0.16 m.

6.4.2.5 *Lead*

The sediments between 2.00 m and 0.92 m in core TCA9b record quite consistent lead concentrations, which fluctuate between 10 and 16 $\mu g\ g^{-1}$. Above 0.92 m, the concentration of lead steadily rises reaching the core's maximum value of 26.7 $\mu g\ g^{-1}$ at 0.76–0.74 m. Values decline above this maximum, with the pattern of lead for the remainder of the core being characterised by a series of well-defined peaks and troughs, with the concentration of Pb lying between 16 and 21 $\mu g\ g^{-1}$. Peaks in Pb in this section of the core were centred on 0.64–0.62 m, 0.52–0.46 m and 0.26–0.24 m. Above 0.16–0.14 m, Pb concentrations increase, with values of over 22 $\mu g\ g^{-1}$ recorded for most of the top 0.10 m.

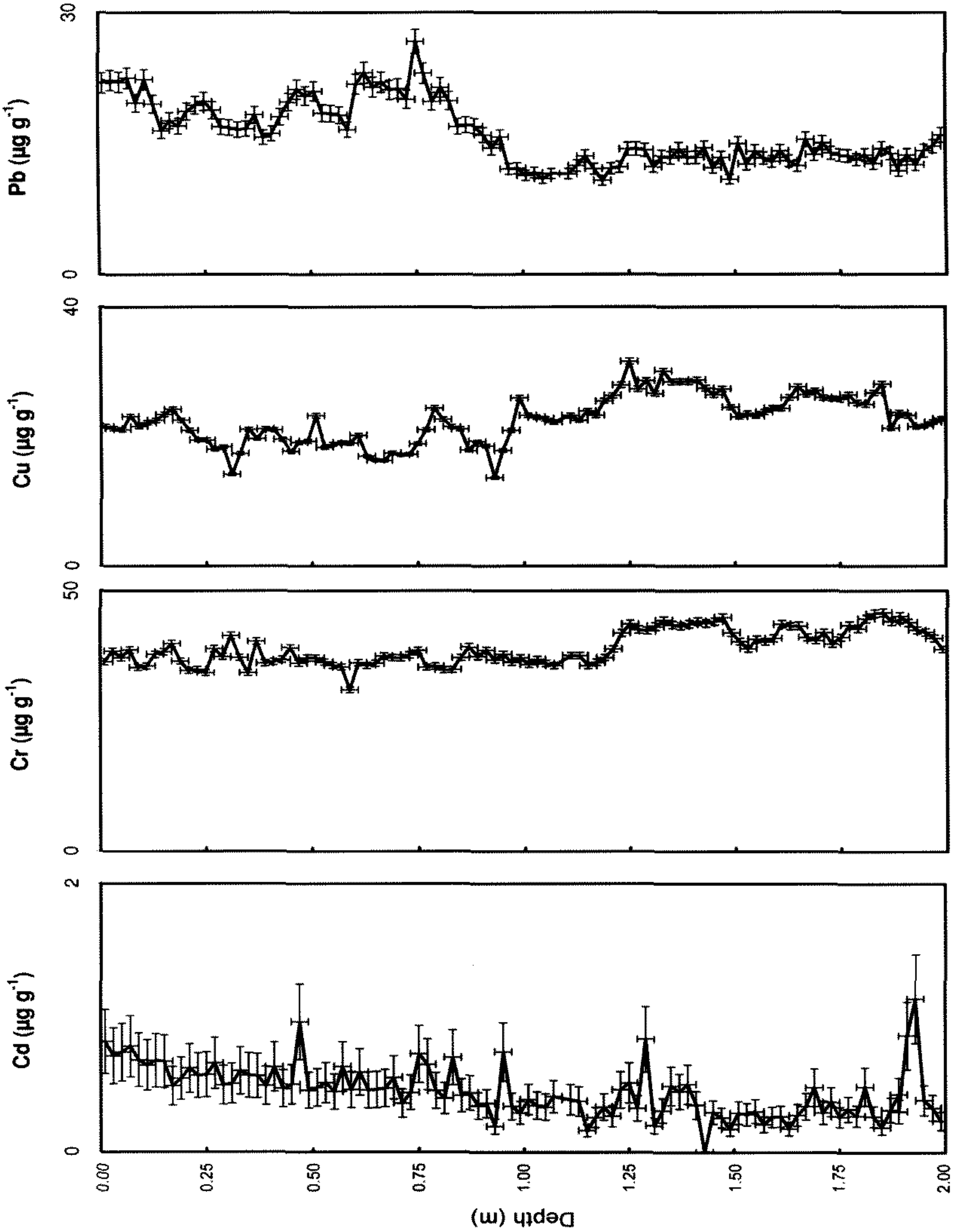


Figure 6.22 Downcore variations in cadmium, chromium, copper and lead, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 95% confidence limits for each measurement.

6.5 Lake sediment palynology

(from Cook et al., in press [analysis by Dr. Dan Penny])

Thirty-eight pollen types were identified in core TCA9b, with an average pollen and spore count of 171 individuals per sample (range = 135–234). The average pollen and spore abundance was 102 919 grains g^{-1} (Figure 6.23). Below 1.26 m, the local (aquatic and littoral) flora is dominated by *Persicaria* (syn. *Polygonum*) and sedges (Cyperaceae). Thereafter, sedges are dominant. Of the dryland herbs, salt-bushes (Chenopodiaceae) are most abundant at 1.36–1.34 m, but decline above this. Grass (Poaceae) is relatively common, though variable, throughout the section, while pollen of the daisy family (Asteraceae) is most strongly represented above 1.16 m. Little obvious change is apparent among the woody taxa, with Casuarinaceae dominant throughout. *Eucalyptus* pollen appears to suffer a decline in abundance above 1.20 m, although the significance of this is difficult to assess without the context of the full pollen sequence (that is, such a decline may be within the amplitude of long-term variability of this genus). No exotic plant pollen is apparent.

The earliest unequivocal palynological evidence for catchment disturbance occurs at a depth of 1.00–0.98 m. The first record of exotic pollen, from the plantain *Plantago lanceolata*, occurs at this depth, while arboreal pollen taxa, particularly those types derived from the dominant canopy trees *Eucalyptus* and *Casuarina*, begin to decline in abundance. The abundance of herbaceous pollen, derived predominantly from grasses, also begins to increase at this depth. The microspores of the aquatic fern *Marsilea drummondii* first appear at this depth, and increase rapidly in abundance thereafter, perhaps suggesting a change in the trophic status of the lagoon linked to increased nutrient flux from the catchment. These data are consistent with the onset of intensive clearance in the catchment.

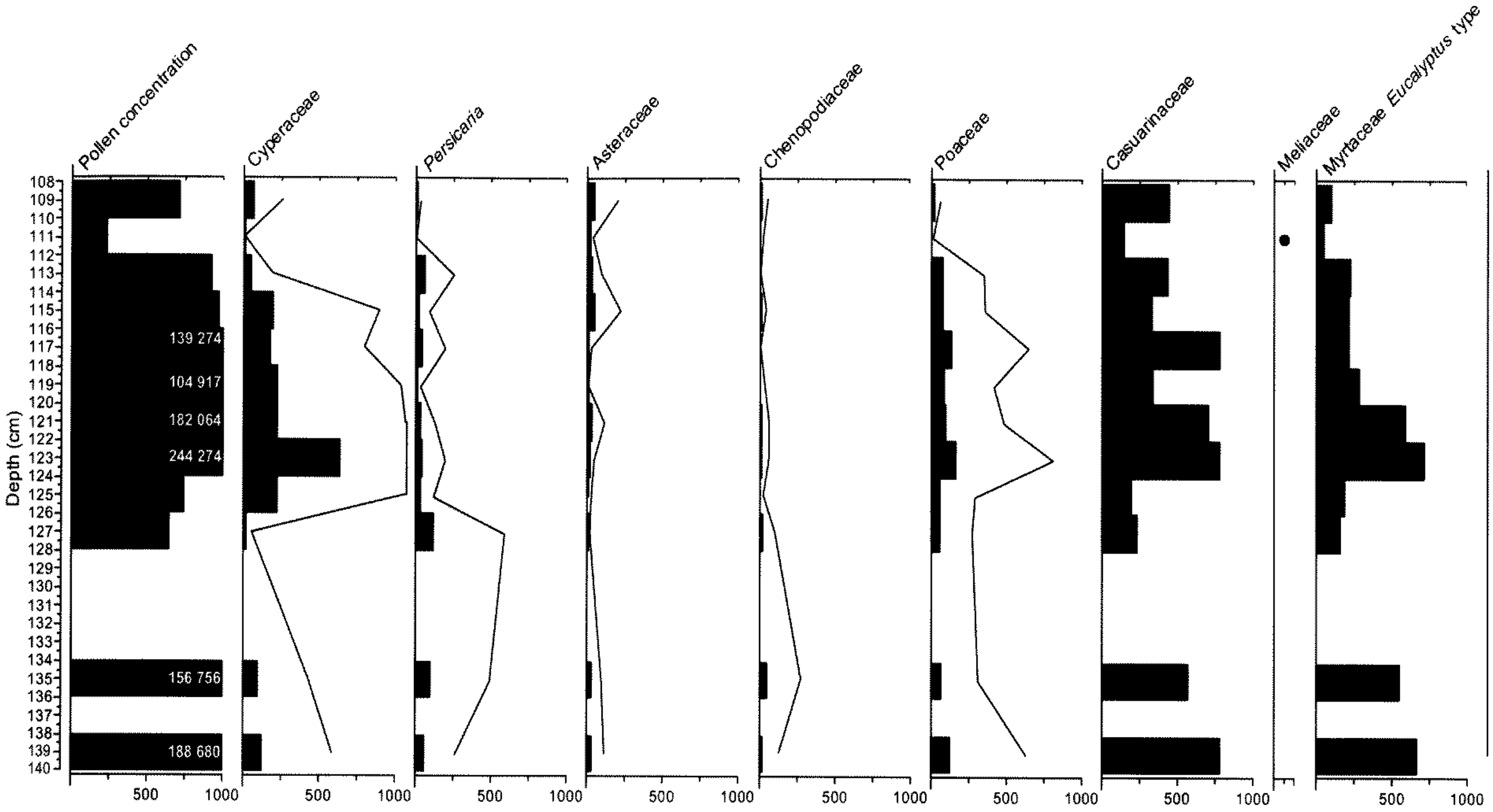


Figure 6.23 The incidence of pollen of selected taxa at depths of 1.40–1.38 m, 1.36–1.34 m and 1.28–1.08 m, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Data are expressed as the number of individuals (x 10^2) per gram dry weight of sediment. Line graphs represent a 5 x exaggeration of the concentration values to emphasise changes over depth in minor taxa that are not clearly graphed at this scale. A datum point is given for Meliaceae, indicating the presence of a single individual at this depth.

6.6 Chronology of core TCA9b

6.6.1 Radiocarbon (^{14}C) chemistry and age determinations

The results of the radiocarbon age determinations for samples of macroscopic charcoal fragments, and microscopic organic extracts from core TCA9b are presented in Table 6.1. Conversion from radiocarbon ages (years BP) to calibrated calendar years (AD) was performed using CalPal (the Cologne Radiocarbon Calibration and Paleoclimate Research Package). Plots of the ^{14}C dates in radiocarbon years and the calibrated calendar years versus sediment depth are presented in Figure 6.24.

The radiocarbon ages generally increase with sediment depth. The single exception comes from the radiocarbon measurement made on pollen and other microscopic organic material extracted from those sediments at 1.24–1.26 m depth, which generated an age of 960 ± 40 years BP (uncalibrated) (Figure 6.24). This is four centuries older than those samples which lie immediately beneath it. The analysis, performed by the AMS laboratory of Beta Analytic Inc. USA, was re-checked and found to be chemically correct (Dr Darden Hood, Director of Beta Analytic Inc. USA, personal communication, 2004). There is no visual evidence in the core of sediment mixing, or disturbance of the sequence. On the contrary, the repeatability of the magnetic susceptibility pattern across the basin shows that the core is unlikely to have been disturbed (Figure 6.2).

Two explanations may be advanced to understand why the date is much older than that expected on the basis of the underlying AMS dates. First, the macroscopic organic matter that was extracted may be significantly older than the deposited sediment, due to older organic material being reworked from catchment soils into the lake. This scenario is all the more likely if those sediments at 1.26–1.24 m depth were deposited during a period of catchment erosion, with older pollen and organic material being transported from the catchment into the lake basin. A similar finding of erroneous radiocarbon ages in recently deposited sediments from a nearby catchment has been reported by Blong and Gillespie (1978). The second explanation is that the sample may have been contaminated at some stage. The ^{14}C geochronology, and the resolution of these conflicting dates is discussed in full in section 7.5.

Table 6.1 Carbon-14 ages and calibrated calendar age determinations made on macroscopic charcoal and microscopic organic extracts from core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

Sample Depth (m)	Lab Code	^{14}C age (^{14}C years)	Error ±1 s (^{14}C years)	Calendar age Cal BP	Error ±1 s (years)	Calendar age cal. BP (68%) min	Calendar age cal. BP (68%) max	Median Calendar age (years AD)	Median Calendar age error (years AD)
1.26–1.24	Beta-190236	960	40	864	53	810	917	1086	53
1.28–1.26	Beta-183139	570	40	586	40	545	626	1364	40
1.28–1.26	Wk-8573	430	70	435	81	354	516	1515	81
1.32–1.30	Wk-7707	730	70	661	64	597	725	1289	64
1.94–1.92	OZE079	2090	60	2065	74	1991	2139	-115	74

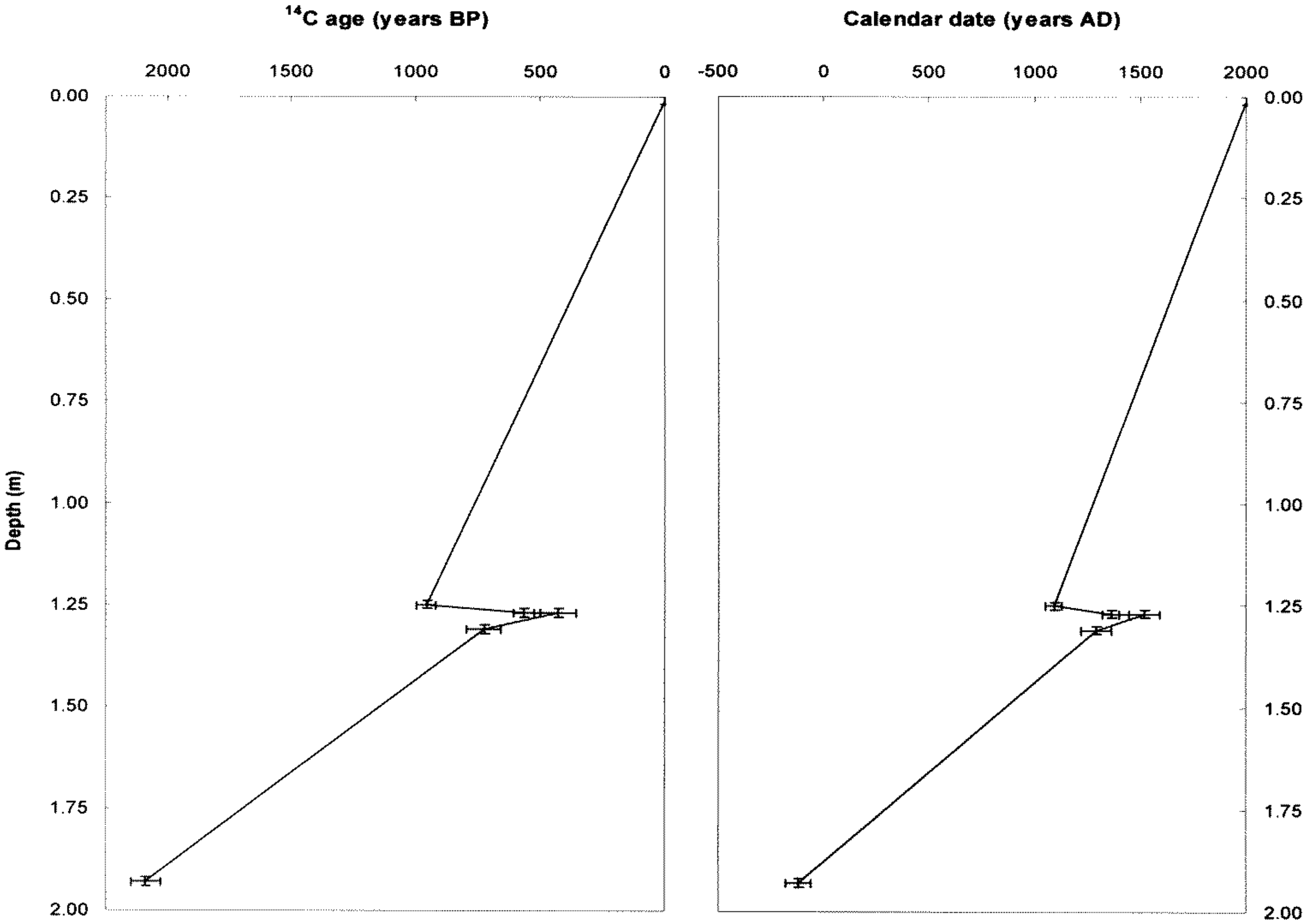


Figure 6.24 Radiocarbon ages and calibrated calendar year dates, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. The curve has been extrapolated to the top of the core for illustrative purposes. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 68% confidence limits for each measurement.

6.6.2 Excess lead-210 activity

The analyses that provided the unsupported ^{210}Pb activities were undertaken in several batches over a three year period. Over such a span there are changes in sample activity, particularly so in the case of those younger samples from the upper part of the sequence. All activities therefore were normalised to the date the core was extracted in 1998 (Appendix 11). The downcore variation in unsupported ^{210}Pb activity in core TCA9b exhibits a broad pattern of exponential decline with increasing sediment depth, superimposed upon which is a series of steps and fluctuations (Figure 6.25). Sediments at 0.88–0.86 m and 0.94–0.92 m record a deficiency in unsupported ^{210}Pb activity.

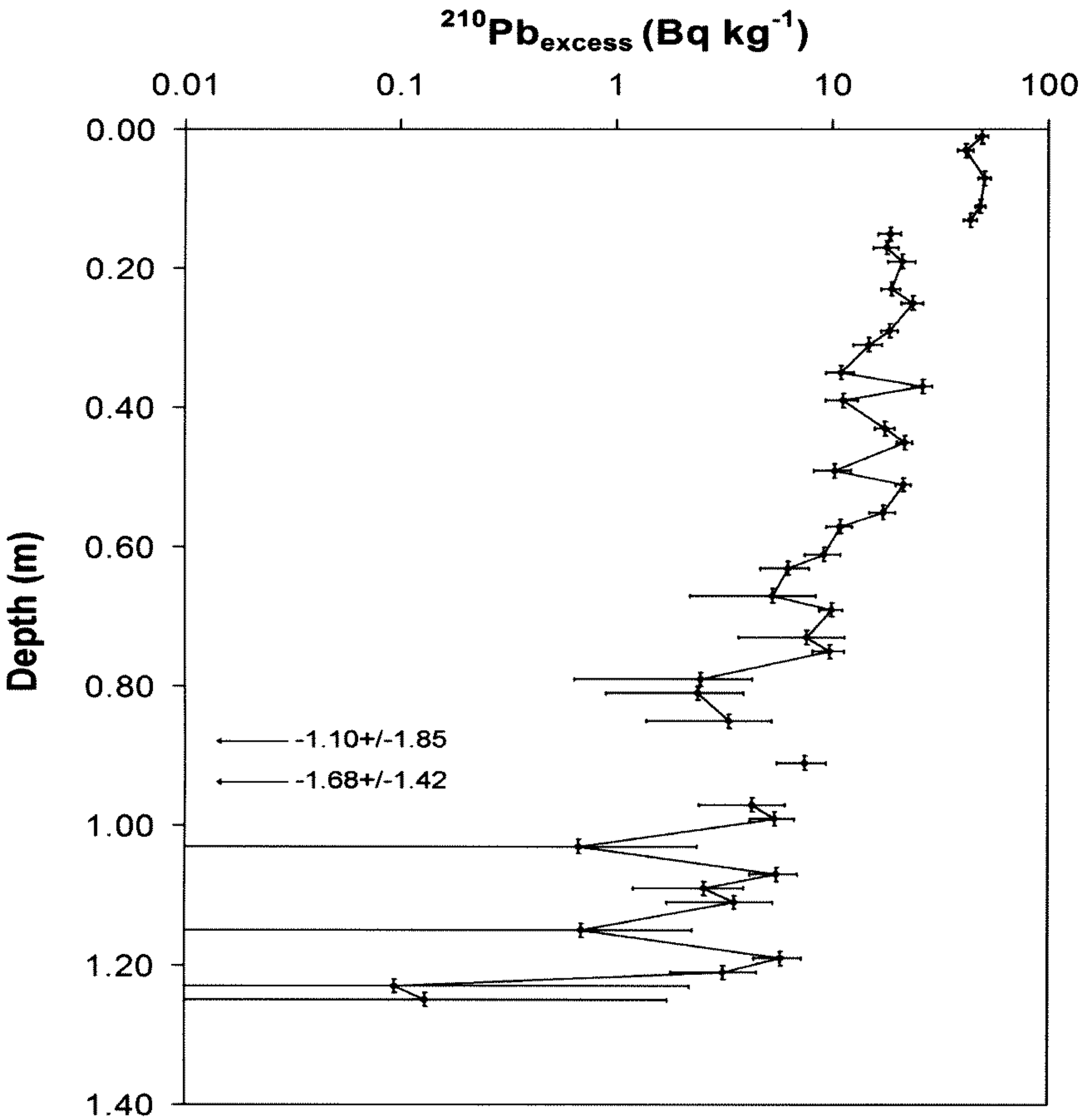


Figure 6.25 Downcore variation in unsupported ^{210}Pb activity, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 66% confidence limits for each measurement. Activities are normalised to the date of sampling in the field.

6.6.3 Caesium-137 activity

The ^{137}Cs activities from core TCA9b are shown in Figure 6.26 (Appendix 12). Below 0.44 m, ^{137}Cs activity is below the limits of detection. Caesium-137 is first detected at 0.44–0.42 m (9.5 ± 1.0 Bq kg^{-1}), and increases in activity upcore reaching maximum values at 0.36–0.34 m (20.5 ± 3.4 Bq kg^{-1}) and 0.26–0.24 m (19.0 ± 2.2 Bq kg^{-1}). From 0.24 m to the top of the core, ^{137}Cs activity decreases rapidly, reaching 3.6 ± 0.3 Bq kg^{-1} at 0.04–0.02 m.

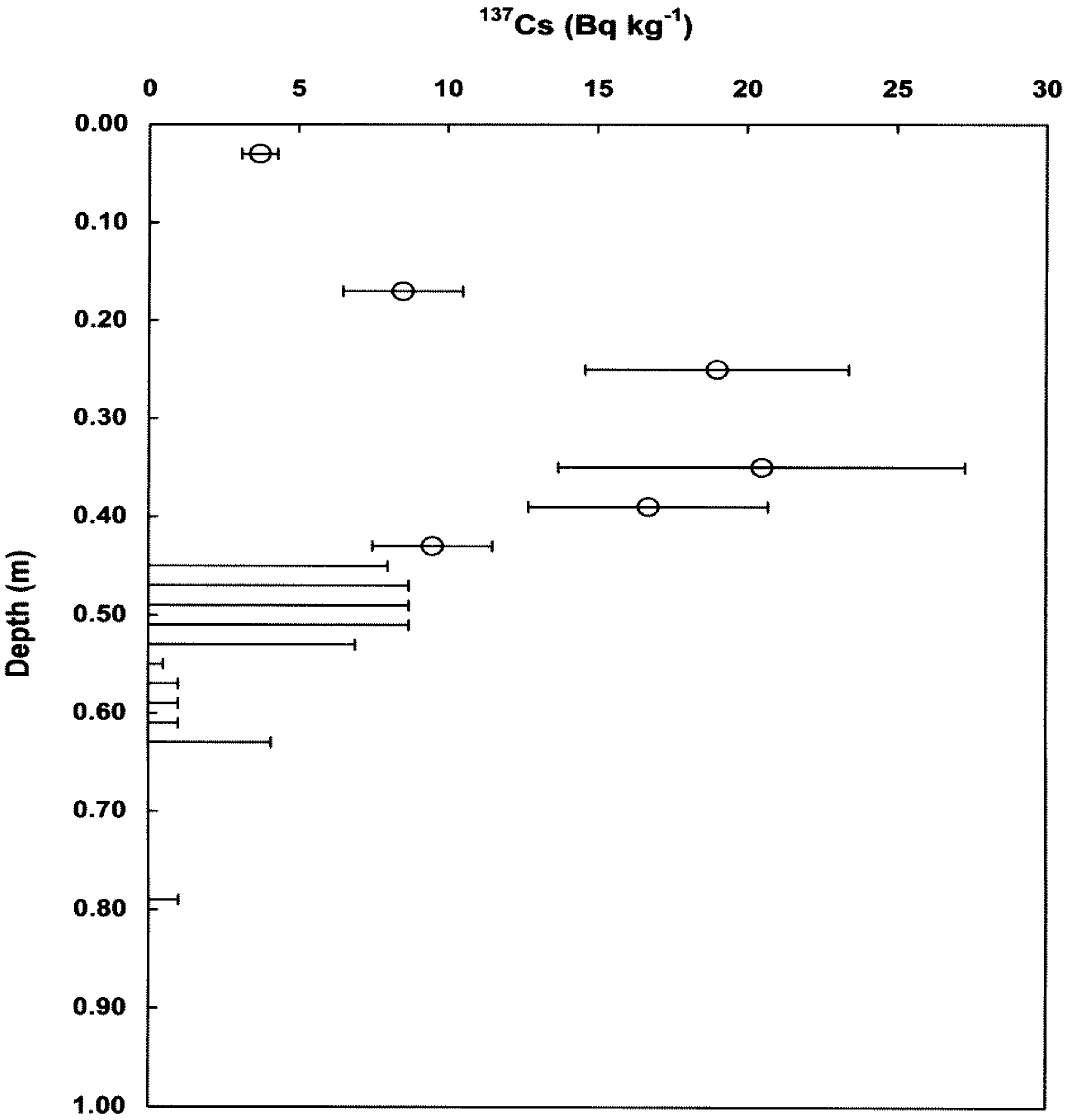


Figure 6.26 Downcore variation in ^{137}Cs activity, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 66% confidence limits for each measurement. The replicate measurement of sample 0.56–0.54 m confirmed no detectable ^{137}Cs activity at this depth.

6.7 Soil core TCWC1

6.7.1 Volume-specific magnetic susceptibility

The downcore measurements of VSMS for soil core TCWC1 are presented in Figure 6.27. From the base of the core to 0.338 m, susceptibility values are low (<50 x 10^{-5} SI units). The top 0.30 m of the soil recorded high values of susceptibility, which fluctuate between 300 and 700 x 10^{-5} SI units. Peaks in susceptibility were recorded at 0.343–0.309 m (552.5 ± 2.9 x 10^{-5} SI units), 0.206–0.172 m (653.5 ± 1.0 x 10^{-5} SI units, the maximum value recorded) and 0.069–0.034 m (559.0 ± 0.0 x 10^{-5} SI units).

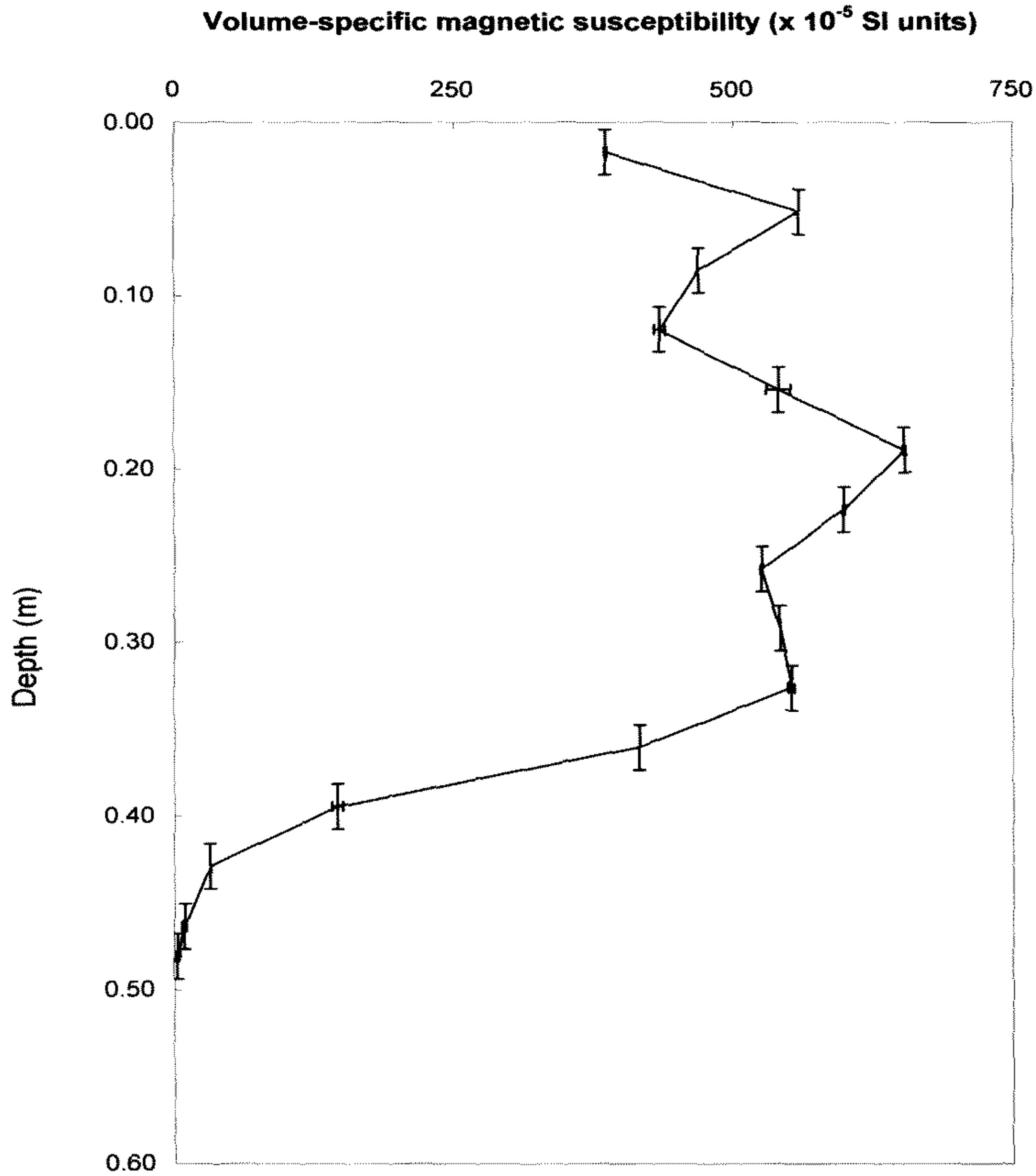


Figure 6.27 Downcore variation in volume-specific magnetic susceptibility, soil core TCWC1, Tocal Homestead Lagoon catchment, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.343 m), while horizontal error bars represent 95% confidence limits for each measurement.

6.7.2 Geochemistry

The results of the ICP-AES analysis of the mud, sand and water-soluble fractions of soil core TCWC1 are tabled in Appendix 13, while the combined water-soluble and mud fraction results are shown in Figure 6.28. Of the 15 elements analysed, the concentrations of two, Mo and As, were lower than the limits of detection of the instrument, and are not considered further. Three main trends in elemental concentrations with depth were recorded (Figure 6.27): little variation with depth, little variation with depth with a large peak value at 0.17–0.14 m or 0.14–0.11 m, and high concentrations at the top of the sequence decreasing downcore.

Concentrations of Fe, Ni and V remain near constant through the soil profile. The concentration of Al is also constant throughout the soil, except for the uppermost sample, where the concentration of Al saturated the ICP-AES, surpassing the limits of detection. The transitional metals Co, Cr and Cu, along with Ba and Mn, record concentrations that vary little downcore, except for a large peak at 0.17–0.14 m. Concentrations of these elements decrease above this to the levels recorded in the lower part of the soil. Finally, concentrations of Ag, Cd, Zn are at their greatest in the upper soil profile, with values decreasing markedly in the top *c.* 0.10 m. Pb displays a similar downcore pattern, but with concentrations increasing below 0.17 m. The lowest soil material exhibits increased Pb concentrations, with the maximum value of 174 $\mu g\ g^{-1}$ recorded at 0.25–0.23 m.

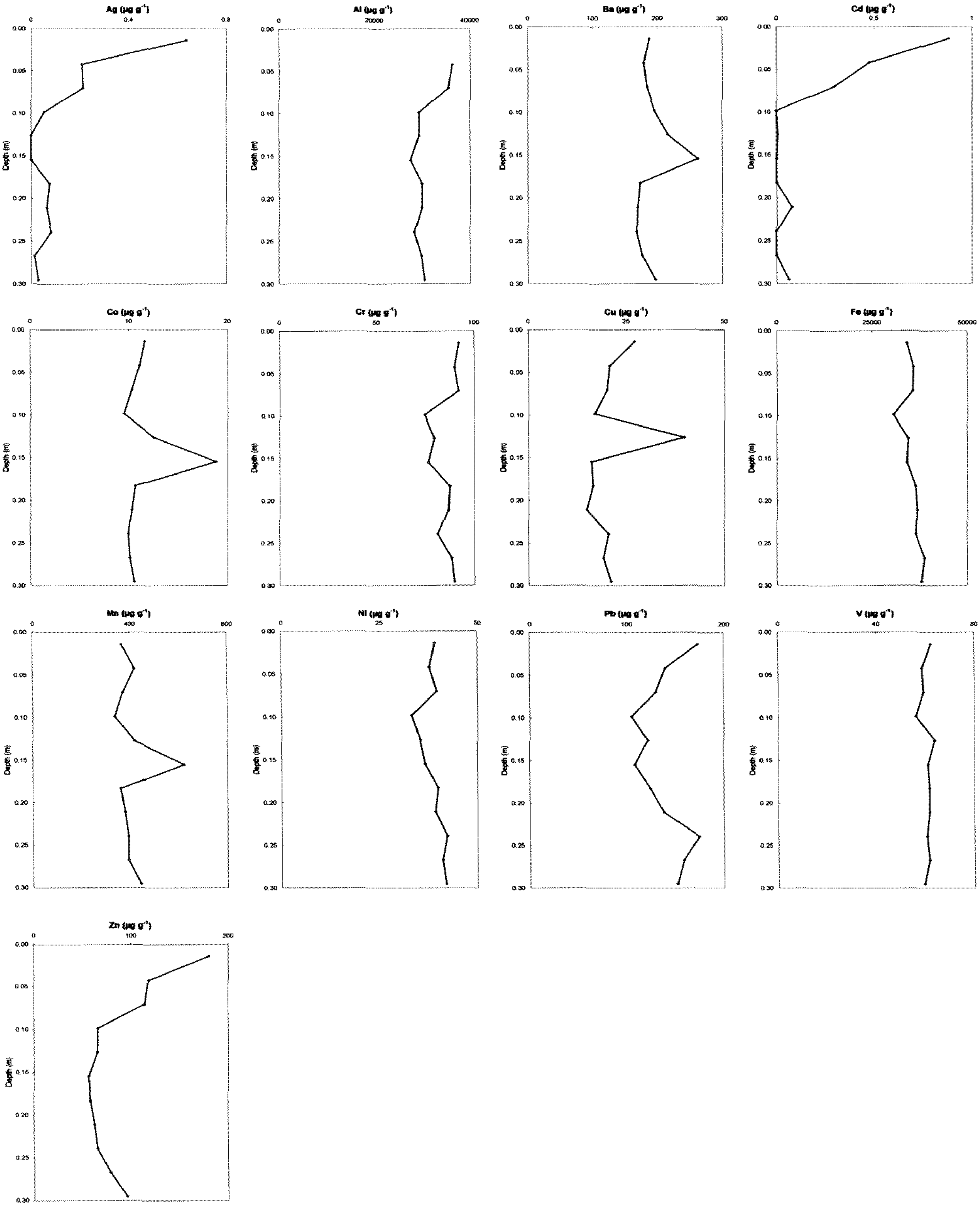


Figure 6.28 Downcore variations in major and minor elemental concentrations of the mud + water-soluble fraction, soil core TCWC1, Tocal Homestead Lagoon catchment, Paterson, eastern Australia. The concentration of Al in the sample from 0.00–0.02 m saturated the spectrometer, and no reading could be made.

6.8 Tocal catchment soils

6.8.1 Geochemistry

The results of the ICP-AES chemical analysis of the 13 catchment soils are tabled in Appendix 14. These include the original analyses of the 13 samples ground by porcelain mortar and pestle (section 5.92) and the 13 samples ground by tungsten carbide disc mill (section 5.9.2) (Figures 6.29 and 6.30).

With the exception of Se and V, all elemental concentrations were above detection limits. The two low-concentration elements will not be considered further. Cd levels, while above detection limits, were low, with sample TS24N recording values below detection limits. Unfortunately, Cd measurement was not undertaken during the replicate analysis of samples T23, T18g and TS18g, so no additional information is available on the repeatability of the Cd measurements.

6.8.2 Data quality

The results of the replicate analyses of samples T23e, T18g and TS18g are presented in Table 6.2, along with the results of the duplicate analysis of sample TS4W (TS4W1 and TS4W2). In the case of samples T18g and TS18g, reproducibility was within 10% for most elements; those elements that exhibited poor reproducibility were typically of very low concentration (e.g., Co). Reproducibility also seemed to decline at the upper detection limits of the ICP–AES, with those elements with concentrations of τ10% exhibiting values between 10 and 20%.

Part of the explanation for the poor reproducibility may have been that the extracts were stored for six months between analyses (albeit at low temperature). This may have resulted in changes in extract concentration due to evaporation and/or sorption to the sample container walls. The repeat analysis of T23e showed very little agreement with any of the original measurements. It is thought that the T23e extract may have been modified after the first analysis. The replicate data will therefore not be considered any further, and the original measurements of T23e will be used alone. In the case of samples T18g and TS18g, mean values will be used for each elemental concentration. The best reproducibility of these experiments was achieved from the two duplicate extractions from soil sample TS4w (Table 6.2); a single mean value for TS4w will be used hereafter.

6.8.3 The effect of grinding procedures on soil chemistry

Samples prepared using the disc mill (the TS series) recorded Al concentrations between 20% and 50% higher than those samples ground using a porcelain mortar and pestle (the T-series) (Appendix 14). Similar differences occurred with Fe, although these were less consistent across the sample group. The mechanical breakdown of clay mineral lattices when using the disc mill may have improved the acid extractant's ability to liberate major elements from the clay minerals into solution. Although elemental concentrations in the TS series were generally higher than those recorded for the porcelain ground set, the difference is typically less than 10% for the remaining elements, and not significant. The T-series results alone are therefore reproduced in Figures 6.28 and 6.29.

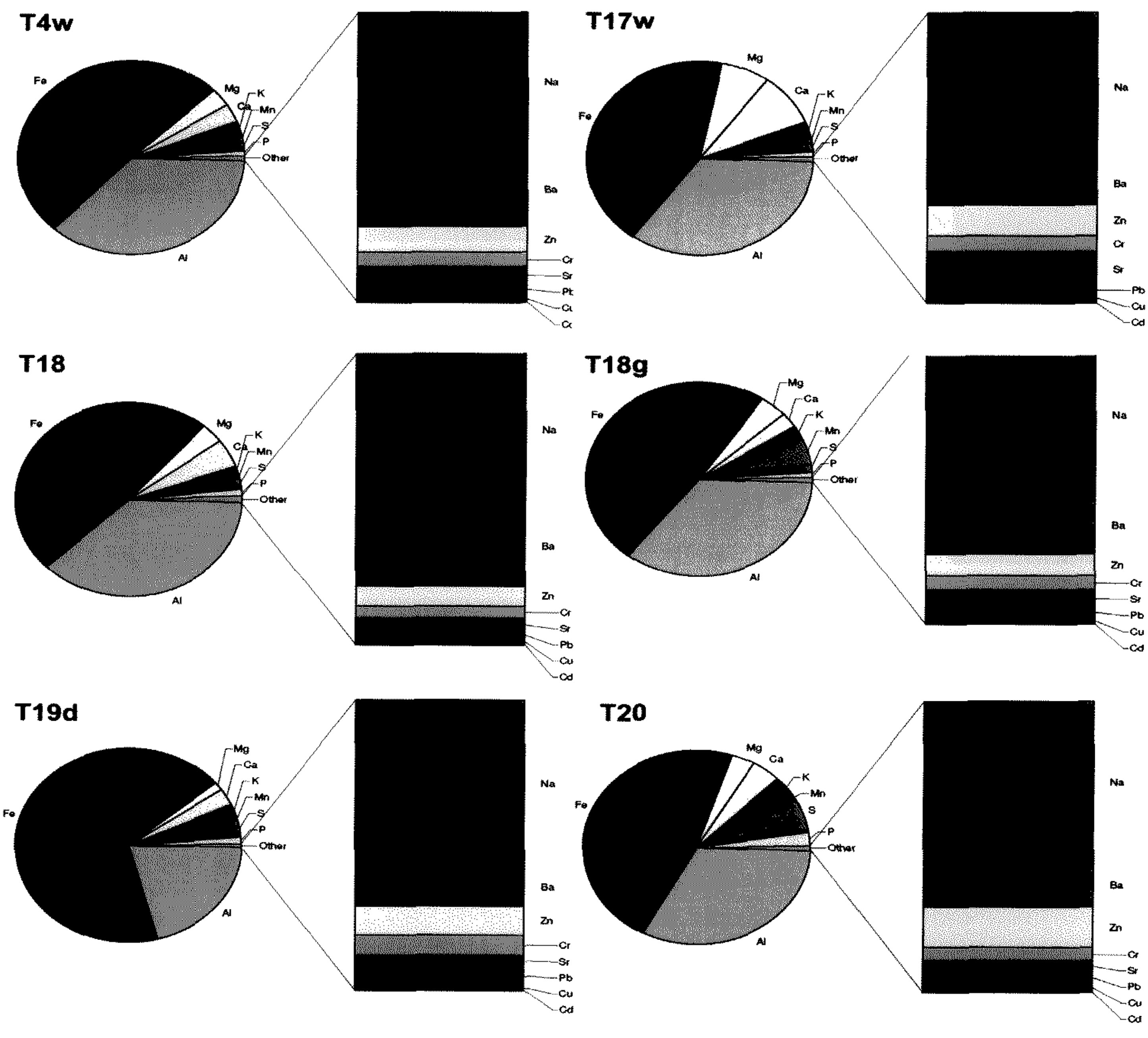


Figure 6.29 The chemical composition of soils T4w–T20 (total fraction), Tocal Homestead Lagoon catchment, Paterson, eastern Australia.

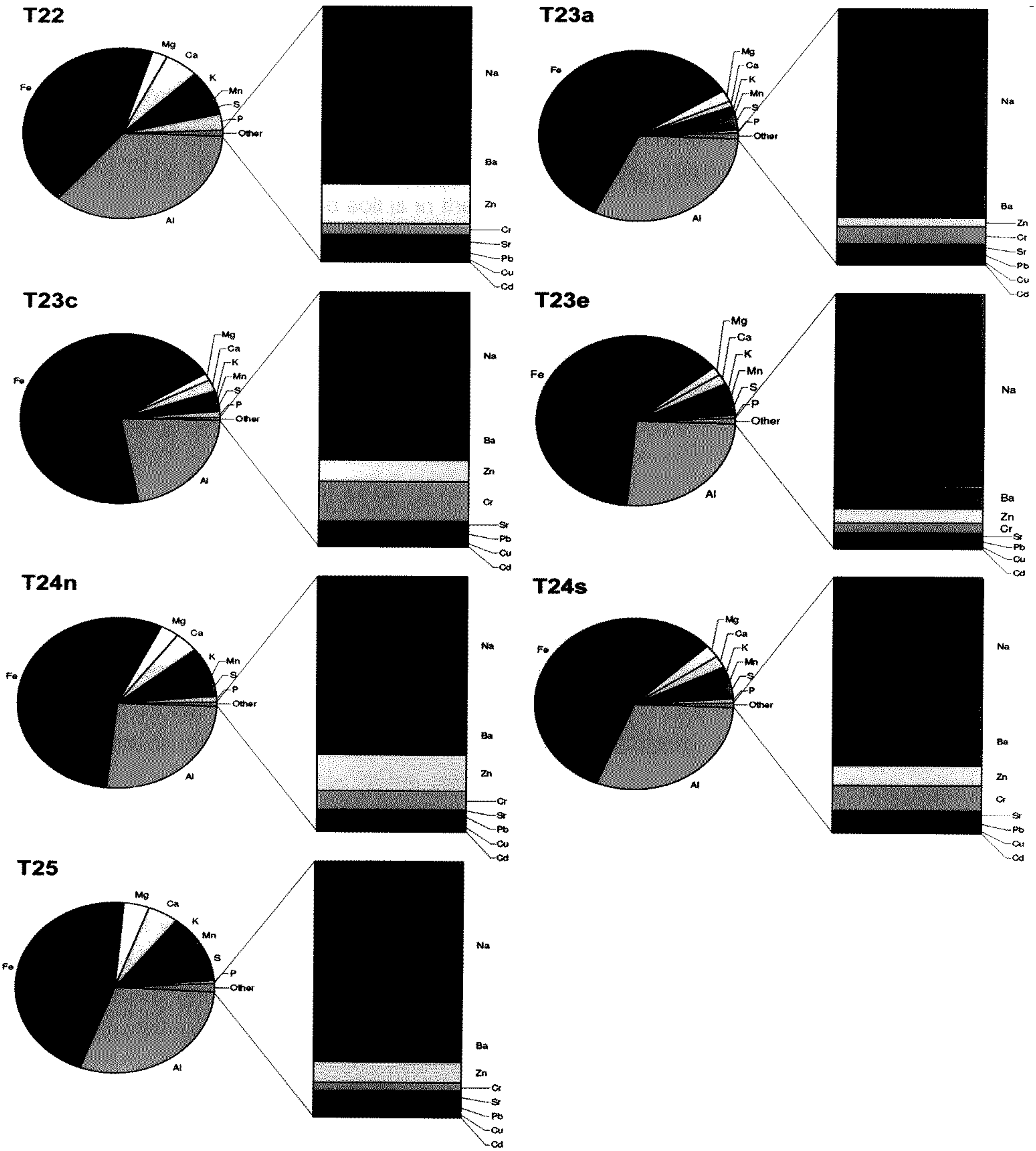


Figure 6.30 The chemical composition of soils T22–T25 (total fraction), Tocal Homestead Lagoon catchment, Paterson, eastern Australia.

6.8.4 Chemistry of soil groups

Considerable variability was recorded between the catchment soils in several of the elements, with Mg, Ca and Sr having 95% confidence limits more than double the mean concentrations of all the soils. The spread of values for these elements may make them useful in chemically identifying different soil types at Tocal. The Mn data also exhibit high levels of variance. However, this is due to the anomalously high Mn content of the 18g soil (samples T18g and TS18g), located on the western slopes of the lagoon. The concentration of Mn in this one soil is in the order of 3000 µg g^{-1}, while the range of values for all other soils at Tocal lies between 34 and 585 µg g^{-1}. In general, the hydrosols and alluvial soils from the low relief areas of the catchment tend to have the greatest elemental concentrations, particularly higher levels of major elements other than Al or Fe (e.g., soil T17w). Gravelly, sandy loam soils sampled from further uphill in the catchment, such as T22, T24s and T24n (Figure 5.6) generally have lower elemental concentrations, possibly reflecting their greater silica content and resistance to acid digestion.

6.8.5 Comparison of core TCA9b, catchment soils and the palaeosol

By characterising the chemistry of the Tocal lake sediments and the potential sediment sources within the catchment, some understanding of the origins of the sediment in Tocal Homestead Lagoon may be made. Focusing on the concentrations of Al, Ba, Cd, Cr, Cu, Fe, Mn, Pb and Zn (those elements studied from the lake sediments, the catchment soils and the palaeosol), some information on the changing source of sediments into the lake over time may be obtained. Elemental concentrations were expressed as ratios to avoid the problems of closed geochemical data sets (Aitchison, 1986). A bivariate plot of Mn/Fe versus Cu/Al clearly separates those lake sediments above 1.26 m from those below (Figure 6.31). Soil samples that lie within the 95% confidence limits of a lake sediment sample are considered to be a potential sediment source (Table 6.3). Ellipses of 95% confidence surrounding the mean of each group show that the basal sediments from the lake are most similar to the total fraction of the buried soil (TCWC1). One catchment soil sample also lies within this group, T25 from the southern slopes of the lake. The larger, second ellipse (the 95% envelope for sediments from 1.26–0.00 m depth) contains several notable groupings within it.

The lower right–hand corner of the post-contact sediments ellipse contains the majority of the mud fraction of the buried soil samples (soil TCWC1). This segment of the ellipse also contains the stratigraphically lowest samples in the 1.26–0.00 m group (from 1.26 m to *c.* 1.00 m depth), which are most similar chemically to the TCWC1 mud fraction samples. The higher sediments from the lake (>0.80 m depth) typically plot in the top half of this ellipse, and cluster alongside some of the catchment samples (n = 6).

Table 6.2 Duplicate chemical analyses of soils from the catchment of Tocal Homestead Lagoon, Paterson, eastern Australia. Where no data ('nd') were available for comparison (e.g., the Mn data for sample TS18g), the single value available is accepted as representative, and tabled in the 'mean' row of data. Values in italics are the results of repeat analyses of diluted extracts. Values in bold denote unacceptable reproducibility (>20%), all of which were recorded for sample T23e. Only the first set of measurements on this sample is considered any further.

	Al	Ba	Ca	Cr	Cu	Fe	K	Mg	Mn	Na	P	Pb	S	Sr	Zn
Sample	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)
T23e (run1)	*6332*	32.7	952	15.8	6.31	*13391*	1281	784	125	114	270	10.6	835	nd	29.5
T23e (run2)	6460	24.7	441	12.0	2.16	13289	750	202	63	251	136	11.5	nd	4.86	22.5
% difference	2.0	**24.6**	**53.7**	**24.0**	**65.8**	0.8	**41.5**	**74.2**	**49.1**	**120**	**49.7**	11.8	nd	nd	**23.6**
Mean	6396	28.7	697	13.9	4.23	13340	1016	493.3	94.0	182	136	11.5	835	4.86	26.0
TS18g (run1)	*31723*	147.5	*2059*	37.6	15.8	*38664*	1722	2745	nd	366	496	24.3	748	41.7	51.2
TS18g (run2)	32327	141.9	2006	38.0	16.7	38148	1734	3097	3116	424	502	nd	nd	42.8	59.0
% difference	1.9	3.8	2.6	1.0	5.0	1.3	0.7	11.4	nd	13.6	1.2	nd	nd	2.5	13.1
Mean	32025	144.7	2033	37.8	16.3	38406	1728	2921	3116	395	499	24.3	748	42.3	55.1
T18g (run1)	*26160*	149.8	2065	34.9	16.4	*36940*	1572	2619	nd	379	482	26.0	1158	42.3	50.6
T18g (run2)	27576	141.5	2122	36.1	17.3	37846	1647	3177	3381	453	500	27.0	nd	42.4	61.2
% difference	5.1	5.5	2.7	3.2	5.1	2.4	4.5	17.6	nd	16.4	3.7	3.4	nd	0.3	17.3
Mean	26868	145.6	2094	35.5	16.9	37393	1609	2898	3381	416	491	26.5	11584	42.3	55.9
TS4w1	24634	108.5	1083	23.9	7.5	nd	1376	1640	343	243	319	16.4	771	24.7	36.8
TS4w2	21291	108.4	1057	23.3	8.1	282045	1325	1592	335	240	305	15.8	785	24.3	35.8
% difference	15.7	0.1	2.4	2.5	6.9	nd	3.9	3.1	2.6	1.5	4.6	3.6	1.9	1.8	2.9
Mean	22963	108.4	1070	23.6	7.8	282045	1350	1616	339	242	312	16.1	778	24.5	36.3

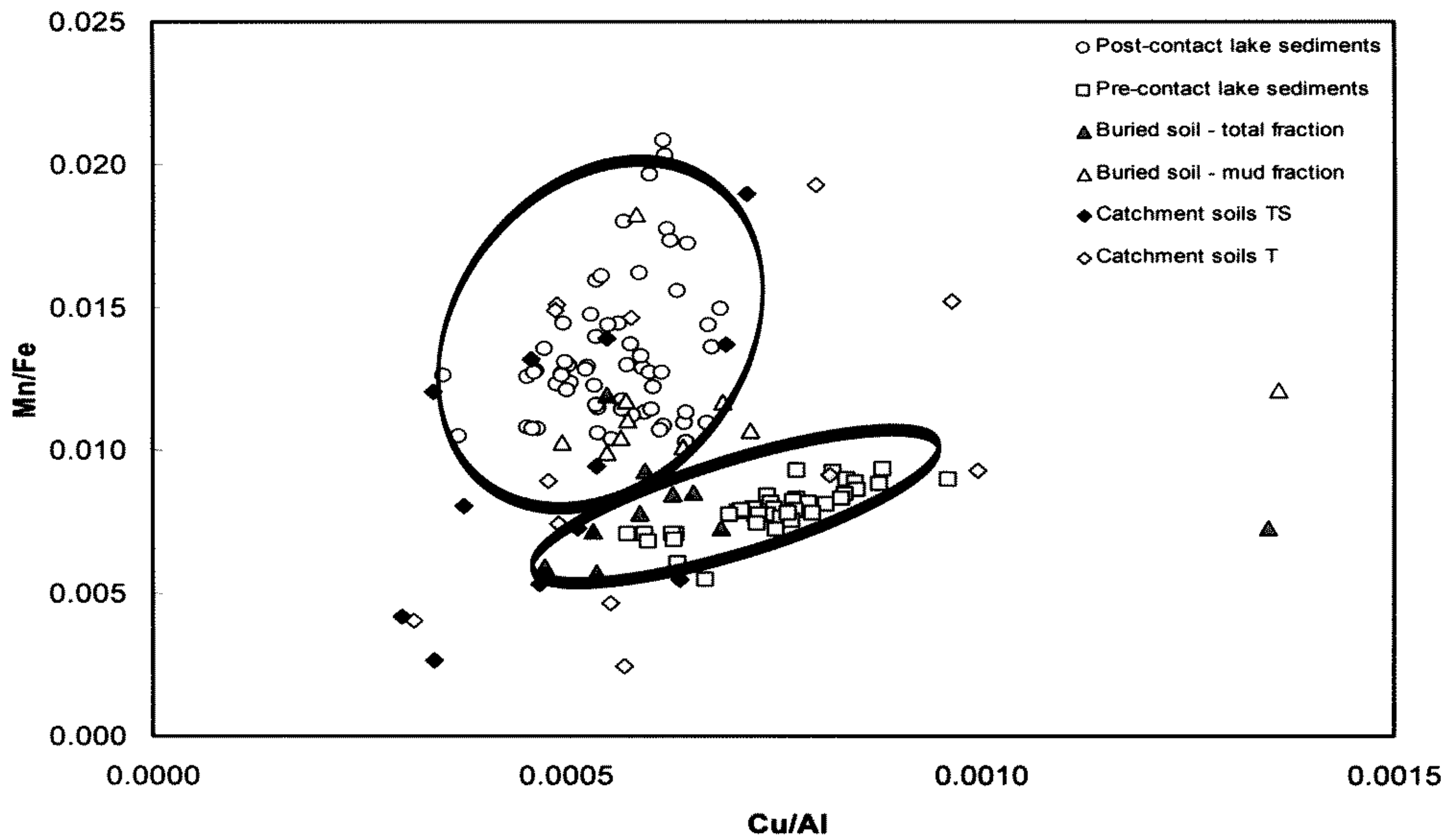


Figure 6.31 The variation of Mn/Fe and Cu/Al data of core TCA9b, 26 soil samples (13 soil profiles, two sets of sub–samples) and two fractions of a palaeosol from the catchment of Tocal Lagoon, Paterson, eastern Australia. Ellipses of 95% confidence surround the mean values of the 2.00–1.26 m and 1.26–0.00 m lake sediment groups.

Half of these catchment samples are from the TS-series; those soil samples prepared differently to the lake sediments, and which recorded elevated Al and Fe concentrations (section 6.8.3). An artefact of this is that the Mn/Fe ratio of these samples is (artificially) quite low compared with the T-series of catchment soil samples. Although geochemical matches can be made between the lake sediments and soils from the TS series (Table 6.3), less confidence is given to these as a result. However, several linkages were established between catchment soil samples T18, T20 and T4w (prepared using the same procedure as the lake sediment samples) and sediments from core TCA9b (Table 6.3).

To demonstrate that these patterns are not simply a function of the analysis of these four elements, principal component analysis (using a covariance matrix), of the entire geochemical data set was undertaken. To do this, preparation of the data involved the exclusion of samples TCWC1.1 and TCWC2.2. Preliminary testing showed that the anomalously high Al values for these samples dominated the principal component analysis, distorting all other patterns. Secondly, the single missing value in the data, the Cd concentration of TS24n, had to be estimated, otherwise this sample could not have been included in the analysis. The mean concentration of Cd of the two closest soil samples from the catchment (TS24s and TS25) was used as the estimate of this missing value. To overcome the problems of closed data sets, elemental concentrations were expressed as their ratios with Al and each ratio was standardised to a mean of zero and unit variance.

The first four principal components explain 92% of the total variance of the data set. The first and second principal components displayed little overlap between the catchment soil, palaeosol and lake sediment data. A biplot of the scores of the second principal component versus the fourth shows some associations similar to those shown produced in the Cu/Al versus Mn/Fe biplot (Figure 6.32), though the relationships are less clear. That some agreement between the two contrasting analyses can be seen is reassuring, considering the greatly increased variability displayed by the entire data set. As with the Cu/Al versus Mn/Fe biplot, the only potential sediment sources to lie within the 95% confidence ellipse of the 2.00–1.26 m sediments are from the total fraction of soil TCWC1. The greatest difference in the multivariate space is that there are no overlaps between the post-contact lake sediments and the buried soil. The only match for the sediments from this post-contact group is with the catchment soil samples.

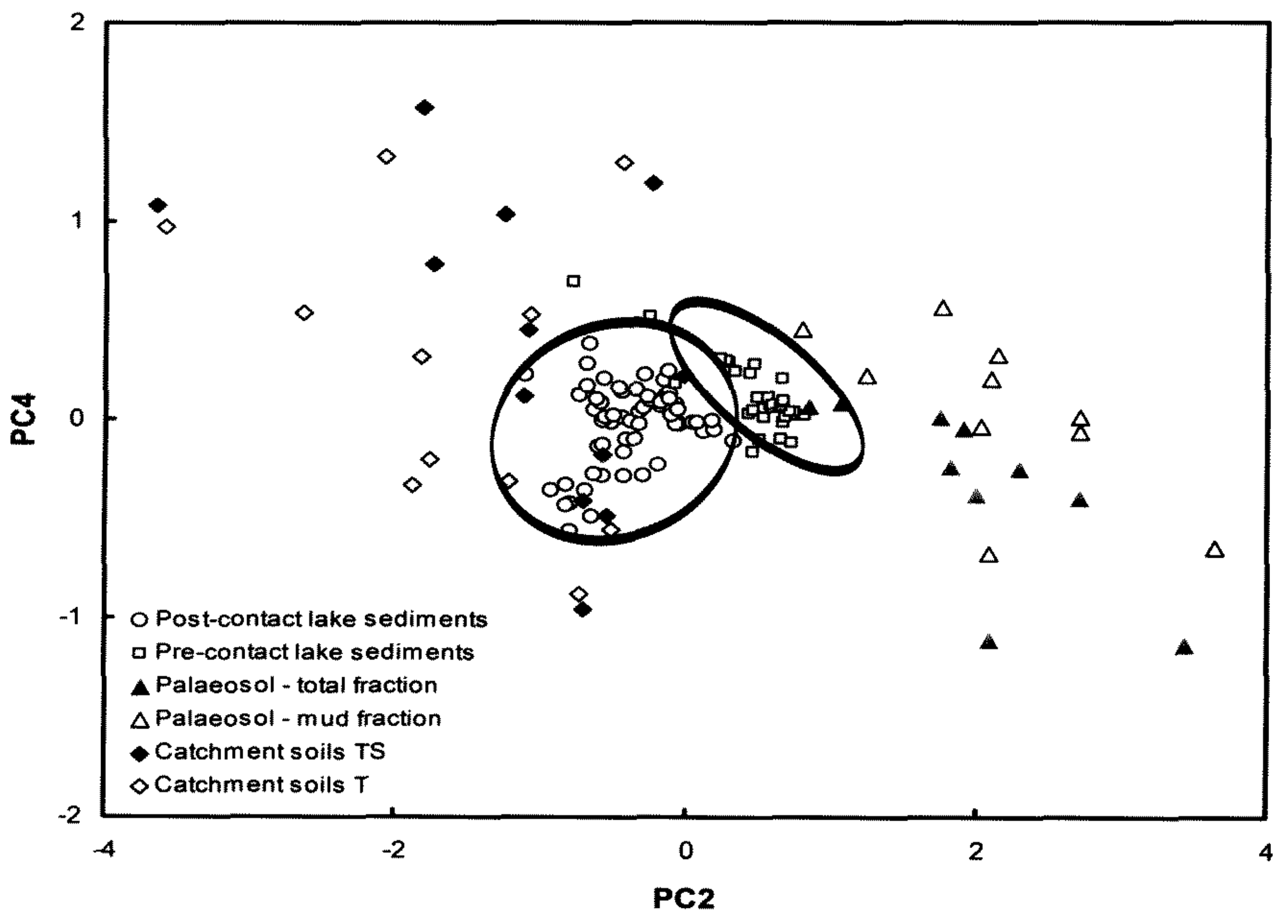


Figure 6.32 Principal component analysis (by covariance) of Ba, Ca, Co, Cr, Cu, Fe, K, Mg, Mn, Na, Ni, P, Pb, Sr and Zn normalised against Al and standardised for each sample from core TCA9b (Tocal Homestead Lagoon), 26 soil samples (13 soils, two sub-samples from each) and two fractions of a palaeosol (TCWC1) from the catchment of Tocal Homestead Lagoon. Ellipses of 95% confidence surround the mean values of the sediments from 2.00–1.26 m and 1.26–0.00 m depths in core TCA9b.

Table 6.3 Provenance of the sediment in Tocal Homestead Lagoon, 1.26 to 0.00 m depth, based on the statistical comparison of the Cu/Al and Mn/Fe content of the sediments from core TCA9b with two fractions of a palaeosol (TCWC1) and 13 soil samples from the lake's catchment. A statistically conservative approach has been taken in assigning sediment sources to avoid spurious connections: sources are connected to sediment depths if the source data lie within ± 5% of the lake sediment's value.

Mean lake sediment depth (m)	Sediment source	Description
0.17	TS20	Catchment soil, TS–series
0.27	TS18	Catchment soil, TS–series
0.31	TS4W	Catchment soil, TS–series
0.33	TS18	Catchment soil, TS–series
0.49	T18	Catchment soil, T–series
0.55	TS24S	Catchment soil, TS–series
0.57	T18	Catchment soil, T–series
0.65	T20, T4w	Catchment soil, T–series
0.67	TS18	Catchment soil, TS–series
0.77	TS24S	Catchment soil, TS–series
0.81	TCWC2.4	Mud fraction of palaeosol
0.83	TCWC2.4	Mud fraction of palaeosol
0.85	TCWC2.3	Mud fraction of palaeosol
1.03	TCWC2.2	Mud fraction of palaeosol
1.05	TCWC2.4	Mud fraction of palaeosol
1.07	TCWC2.4	Mud fraction of palaeosol
1.11	TCWC2.3	Mud fraction of palaeosol
1.15	TCWC2.4	Mud fraction of palaeosol
1.19	TCWC2.11	Mud fraction of palaeosol
1.23	TCWC2.10	Mud fraction of palaeosol

6.9 Soil aggregate stability

The results of the Emerson aggregate test (EAT) for soils TCWC1 and TCS1 (a catchment soil sampled by Cook [1998] from the western slope of the lake) are presented in Table 6.4, along with the corresponding LOI_{430} data. Emerson aggregate stability classes for the palaeosol TCWC1 are very low, ranging from values of 1 at the top of the profile to 3 at 0.05 m and 0.15 m. Below 0.19 m, EAT classes of 2 were recorded for the remainder of the profile. EAT classes for the modern catchment soil TCS1 were significantly higher; the topsoil recorded a class of 6, while the subsoil material was of class 5 stability. The low aggregate stabilities of the upper material in soil TCWC1 are exceptional considering the relatively high LOI_{430} (%) values for those samples, a maximum of 10.7% for the top sample.

The percentage of aggregates (2 to 0.5 mm) retained after wet sieving (Aggregate Stability) for each soil demonstrates a similar trend. Aggregate Stability values for TCWC1 range from 5.5% at the soil's surface to a maximum of 11.8% at 0.36 m. It is important to note that soil TCWC1 was weakly aggregated and obtaining enough aggregates for the measurements was difficult. Those aggregates wet–sieved were generally very unstable, and once again organic matter content (LOI_{403}) appears to have little influence. Aggregate Stability values for the topsoil and subsoil sampled from TCS1 were much higher, with the topsoil recording 63.6%, an order of magnitude greater than some values recorded for TCWC1. The subsoil of TCS1 generally had fewer stable aggregates (and less aggregate material) with 22.6% of the 0.5–2.0 mm sized aggregates enduring the wet sieving process.

Table 6.4 Soil aggregate stability in soils TCWC1 and TCS1 from the catchment of Tocal Homestead Lagoon, Paterson, eastern Australia.

TCWC1				TCS1			
Mean depth (m)	LOI_{430} (%)	EAT Class	Aggregate Stability (%)	Soil type	LOI_{430} (%)	EAT Class	Aggregate Stability (%)
0.02	10.7	2	5.5	Topsoil	10.91	6	63.6
0.05	10.2	1	Nd	Subsoil	3.27	5	22.6
0.09	9.3	1	6.0				
0.12	9.3	2	10.5				
0.15	8.9	3	10.7				
0.19	7.4	2	9.7				
0.22	7.7	2	9.4				
0.26	7.6	2	9.8				
0.29	6.8	2	9.6				
0.33	6.9	2	9.4				
0.36	7.3	2	11.7				
0.39	6.1	2	10.8				
0.43	6.0	2	nd				
0.46	6.1	2	nd				
0.50	6.1	nd	nd				
0.53	6.1	nd	nd				

6.10 Catchment drainage density

The results of the analysis of channel lengths and drainage density from historical aerial photographs of the Tocal catchment are presented in Table 6.5. Drainage density at Tocal has generally decreased over the last *c.* 50 years, with peak channel incision recorded in 1967 (Figure 6.33). The most recent measurement, from the aerial photograph of 1996, shows a catchment drainage density of 40 m ha^{-1}, while values have been as high 50 m ha^{-1} (1967).

Table 6.5 The ratio of total channel length and catchment area (drainage density) for the catchment of Tocal Homestead Lagoon, Paterson, eastern Australia, measured from 1958, 1967, 1984, 1993 and 1996 aerial photographs.

Year	**1958**	**1967**	**1984**	**1993**	**1996**
Total length of channels (m)	4301	4626	4004	3832	3749
Catchment area (ha)	93	93	93	93	93
Drainage density (m ha^{-1})	46	50	43	42	40

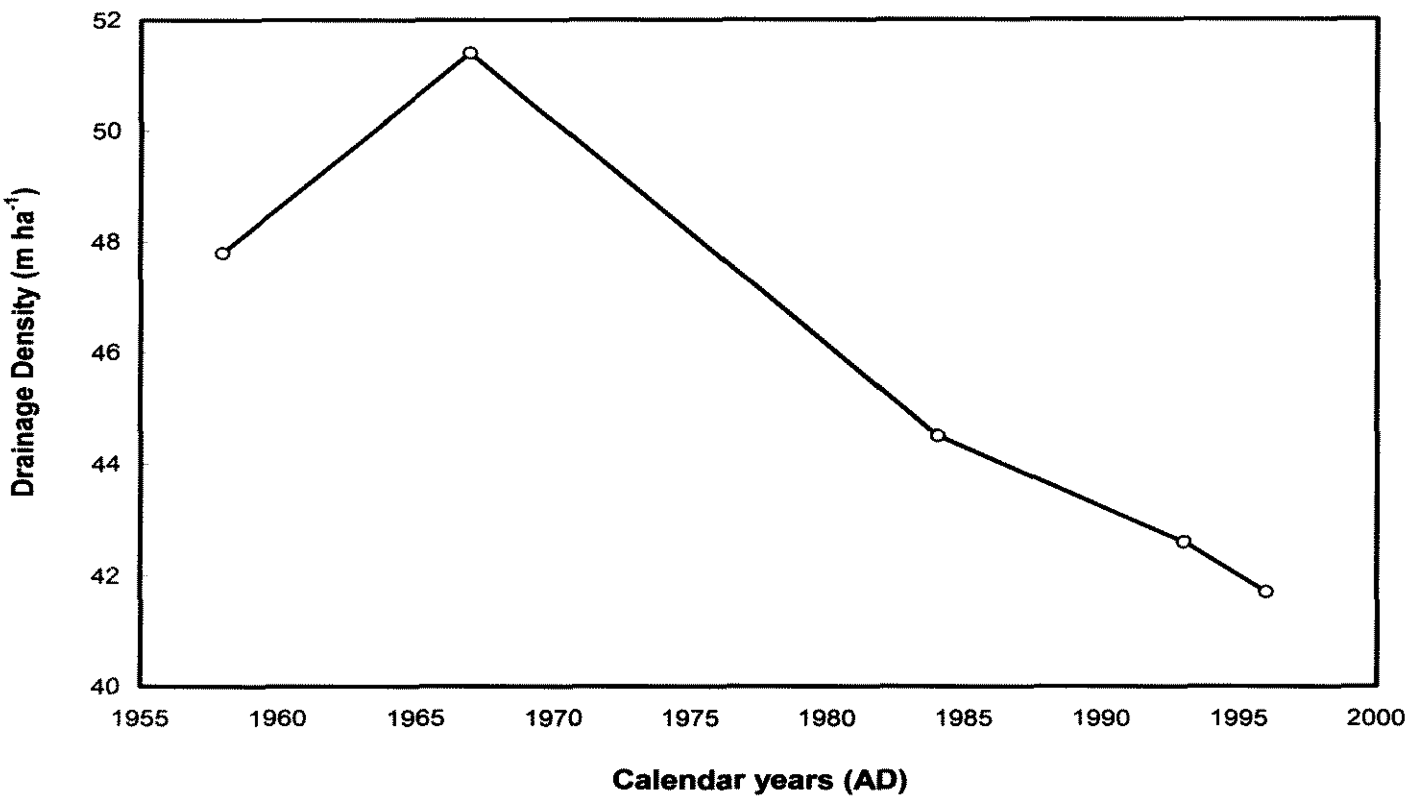


Figure 6.33 Historic changes in catchment drainage density, Tocal Homestead Lagoon, Paterson, eastern Australia.

6.11 Discussion of results

6.11.1 The mineral magnetic record of Tocal Homestead Lagoon

6.11.1.1 Magnetic mineralogy of the sediments: implications for soil erosion

Although significant variability was recorded in each magnetic property measured on the sediments from the master core (section 6.2), most of these properties distinguish between those sediments in the top *c.* 1.25 m of the core and those in the lower part of the core. In the case of χ_{lf}, for example, values rise significantly above the lower, background levels at a depth of 1.26–1.24 m, systematically increasing through three orders of magnitude. The top 1.00 m of the sediments has average χ_{lf} values of *c.* 70 x 10^{-6} m^3kg^{-1}, with a maximum value of nearly 370 x 10^{-6} m^3kg^{-1}. In comparing these data to previously published records from sites worldwide (Dearing, 1999a; Oldfield, 1999), the sediments have a magnetic susceptibility typical of that dominated by ferrimagnetic minerals. In soils and sediments, ferrimagnetic minerals are predominantly iron oxides; magnetite, maghaemite, titanomagnetite and titanomaghaematite (Gale and Hoare, 1991). Magnetite and maghaemite are by far the most common of these iron oxides in environmental samples (Thompson and Oldfield, 1986). Following Maher *et al.*'s (1999, p. 42) summary of diagnostic magnetic properties, the uppermost sediments from the Tocal sequence are interpreted as dominated by magnetite. The S-ratio plot for core TCA9b supports this (Figure 6.14); the sediments deposited in the upper *c.* 1.25 m are most probably composed of magnetically 'soft' minerals (S-ratio approaching 1), indicative of magnetite. As the majority of magnetite is completely magnetised in field strengths of 100 mT (only the finest magnetite is not [Walden, 1999b]), an S-ratio ($IRM_{100\ mT}$/SIRM) approaching 1 indicates a dominance of coarse-grained magnetite. In conclusion, the sediments in core TCA9b above *c.* 125 m are most likely dominated by coarse magnetites.

The type of magnetic mineral in sediments is not the only control on bulk magnetic susceptibility. The dominant factor controlling the magnetic susceptibility of materials within the ferrimagnetic class is the concentration of these minerals in the sample (Dearing, 1999). Other factors that may influence susceptibility (in decreasing order of importance) are magnetic grain size, magnetic grain shape (Thompson and Oldfield, 1986) and inter-grain interactions (Lees, 1997).

6.11.1.2 The provenance of magnetic minerals in core TCA9b, 1.26–0.00 m

The substrate in the catchment of Tocal Homestead Lagoon is composed predominantly of alluvium, conglomerate and some lenses of sandstone (section 2.1), and is magnetically weak. Of 230 sandstones reported by Gale and Hoare (1991, p. 208), the mean χ_{lf} was 18×10^{-8} m^{3}kg^{-1} with a maximum reported value of 948×10^{-8} m^{3}kg^{-1}. Conglomerates were magnetically weaker again, with a mean mass susceptibility of 8.6×10^{-8} m^{3}kg^{-1} and a maximum value of 14×10^{-8} m^{3}kg^{-1} (n = 6). No direct measurements have been made on the parent material at Tocal. Adopting the assumption that the magnetic susceptibility of these materials is similar to the mean values reported by Gale and Hoare (1991), the magnetic susceptibility of the catchment soils of Tocal Homestead Lagoon is up to two orders of magnitude greater than that of their parent materials. Where lake sediments with strong magnetic mineralogy have been derived from weakly magnetic bedrock, the source of magnetite may be the secondary minerals formed *in situ* in soils that have since been eroded (Le Borgne, 1955; Thompson and Oldfield, 1986). Industrial pollutants and atmospheric fallout may also magnetically enhance weakly magnetic soils (Maher, 1998). Atmospheric deposition may be an additional control on magnetic mineralogy in systems where the supply of ferrimagnetic and antiferromagnetic minerals is limited (Gale and Hoare, 1991). However, with Tocal's great distance from any significant pollution sources, and elevated susceptibility values dating from the early 19th century in the Tocal sediments, pollution-based enhancement of soils or sediments at Tocal is not thought to be of particular importance.

Since this is the case, the downcore pattern of magnetic susceptibility in the sedimentary sequence is likely to record the flux of soil-derived magnetic minerals from the catchment through time, and may therefore be interpreted as a record of changes in catchment erosion. Greater volumes of denser magnetic minerals such as magnetite are eroded and transported during relatively higher-energy events (e.g., Thompson *et al.*, 1980). Therefore peak values of magnetic susceptibility may record episodes of denser/coarser magnetic material being liberated from the catchment or greater concentrations/volumes of magnetic minerals being deposited, or both. In either case, the peaks in magnetic susceptibility record instances of enhanced catchment erosion, while episodes of lower susceptibility may record catchment stability.

A final test of whether the Tocal magnetic record is controlled mainly by catchment erosion is to examine the pollen record from core TCA9b (Cook *et al.*, in press) (Figure 6.23, section 6.5) for any connections to the peaks exhibited in the magnetic susceptibility record (Figure 6.8, section 6.2.6). Previous magnetic studies of catchment erosion and lake sedimentation have often found some correlation between peaks in lake sediment magnetic susceptibility and shifts in pollen (the first to observe such a pattern were Thompson *et al.*, [1975]). The synchronicity of both indicators suggests the magnetic peaks reflect periods of increased erosion, with accompanying changes in the amount of grass pollen eroded into the lake. The most convincing relationship between the two measures can be seen above *c.* 1.26 m in the master core, where sharply increasing magnetic susceptibility is accompanied by an increase in the concentrations of Cyperaceae and, to a lesser extent, Poaceae pollen. This is followed by a decline in *Eucalyptus* pollen above 1.20 m that may record the shift in vegetation and the corresponding soil erosion that followed the onset of European agriculture in the catchment in *c.* 1820. More recent peaks in susceptibility seem to be independent of any major shifts in the pollen assemblage, which may be as expected if a major vegetation clearance took place following European settlement. Subsequent periods of increased catchment erosion, whether in response to land-use changes or climatic events, would not necessarily involve any further vegetation modification.

A catchment-erosion interpretation alone, however, cannot account for any modification of magnetic minerals that may take place following deposition in the lake basin. Although many authors have recorded the preservation of lake sediment magnetic mineralogy following deposition (Rummery *et al.*, 1979; Oldfield, 1991; Eriksson and Sandgren, 1999), such findings are not applicable to all lake systems. When sediments possess magnetically weak mineralogy, the detrital record may be obscured by the development of *in situ* ferrimagnets (typically much weaker than magnetite/maghaemite) (Dearing, 1999b), bacterial biomineralization (Van der Post *et al.*, 1997) and magnetic dissolution.

Post-depositional growth of magnetosomes is typically detected by either the microscopic (Hesse and Stolz, 1999) or magnetic (Eriksson and Sandgren, 1999; Oldfield, 1999) examination of lake sediments. Bacterial magnetite consists of long chains of fine-grained (*c.* 0.05 µm) magnetic crystals, which are characterised by very strong ARMs (or particularly high ARM/SIRM or ARM/χ_{lf} ratios) (Maher *et al.*, 1999). Non-biological fine-grained magnetite may be distinguished from magnetosomes using the squareness of hysteresis loops (Oldfield, 1994; 1999; Maher *et al.*, 1999). These loops are based on the relationship between the strength of an applied magnetic field and the intensity of the magnetisation induced in the substance by this field. Squareness of the hysteresis loop is characteristic of the ease of magnetisation of a sample, and indicated by the SIRM/χ_{lf} or ARM/χ_{lf} ratios (Maher *et al.*, 1999). Sediments affected by magnetosome bacteria are typically characterised by values of χ_{ARM}/SIRM τ2 x 10^{-3} m A^{-1} (Oldfield, 1999), which suggests that their presence in the sediments is very unlikely (Figure 6.13). The χ_{ARM}/SIRM ratio shows the Tocal sediments to be several orders of magnitude lower than measurements made on magnetosome-bearing lake sediments in previous studies (e.g., Oldfield, 1999). The χ_{ARM} and LOI_{430} values of the Tocal sediments are also well below those considered by Paasche *et al.* (2004) to characterise magnetotactic bacteria.

Dissolution is a diagenetic process that involves the reduction of fine magnetic minerals. Although chemical reduction by dissolved sulphide has been assumed to be the principal process, bacteria may reduce and dissolve magnetite in the presence of organic matter (Kostka and Nealson, 1995). If the sedimentation rate is low, the dissolution of magnetite at the sediment–water interface may occur simultaneously with the decomposition of organic compounds, especially if anoxic conditions prevail (Canfield and Berner, 1987; Snowball, 1993). Ferrimagnetic grains that lie close to the super paramagnetic–single domain (SP–SD) boundary, which give rise to high χ_{fd} values in sediments, are among the first to dissolve (Karlin, 1990; Leslie *et al.*, 1990). As such, the magnitude of χ_{fd} values has been used to as evidence of magnetic dissolution (Eriksson and Sandgren, 1999). Evidence of magnetic dissolution in sediments may indicate periods of anoxic and reducing conditions, such as those experienced during droughts (Caitcheon, 1998).

Dissolution processes would effectively decrease the concentration of magnetic minerals, resulting in a reduction of susceptibility independent of any changes in sediment flux. Rates of minerogenic sediment accumulation at Tocal Homestead Lagoon are high in the upper 1.26 m of the sequence (section 9.2), while the organic content of the sediments rarely exceeds 10% (section 6.2.4). Post-depositional modification of magnetic mineralogy by dissolution is therefore unlikely in the upper sediments of core TCA9b. However, the lower magnetic susceptibility and typical absence of frequency-dependant magnetic grains below 1.26 m means that some partial magnetic dissolution may have taken place in these gleyed sediments, perhaps sufficient to modify the finest magnetic grains (represented by χ_{fd} values). An alternative explanation is that lower magnetic susceptibilities and zero or low χ_{fd} values in the sediments below 1.26 m may represent a period of lower input of pedogenic ferrimagnetic material into the basin. The low sedimentation rates recorded for the pre-European period (section 9.2) provide ample support for this argument. Susceptibility measurements alone, however, are incapable of resolving which process may be responsible.

6.11.1.3 Interpretation of the S-ratio

The S-ratio (Figure 6.13) records shifts in the relative contribution of paramagnetic and ferrimagnetic minerals in sediments (Dearing *et al.*, 1998). Lower S-ratios may therefore record either a decreased ferrimagnetic mineral content in the sediments or an increased paramagnetic mineral content. It may be possible to choose between these two explanations by checking for parallel increases in magnetically hard paramagnetic minerals such as haematite or goethite using the SIRM-$IRM_{100\,mT}$ plot (Figure 6.13). Below 1.26 m, major shifts in the S-ratio suggest that the relative contribution of magnetically hard minerals varied significantly. Lower S-ratios are recorded for the entire pre-European sequence, with particularly low values from *c.* 1.52 m to *c.* 1.34 m. The SIRM-$IRM_{100\,mT}$ plot shows consistently low values for most of the sediment below 1.26 m, suggesting that decreased S-ratios cannot be explained by an increase in magnetically harder material, which may be either post-depositional in origin, or due to a shift in sediment source to paramagnetic-dominated soil. The only remaining explanation is that the pre-1.26 m sediments are characterised by significantly lower concentrations of pedogenic ferrimagnetic minerals. This is most probably explained by the low flux of material from the catchment, suggestive of stable environmental conditions for much of this period, although post-depositional dissolution of the sediment is also a possibility.

In summary, the magnetic susceptibility record from the lake appears primarily to record changes in the concentration of ferrimagnetic minerals that have been eroded into the basin from the soils of the catchment. These data are thought to reflect the changing volumes of soil-derived materials that have been eroded, transported and deposited in the lake basin from the surrounding catchment.

6.11.1.4 The provenance of magnetic minerals in core TCA9b, 2.00 –1.26 m

The materials deposited in Tocal Homestead lagoon from 2.00 m to 1.26 m (core TCA9b) are characterised by low χ_{lf}, χ_{ARM}, $IRM_{100\,mT}$ and SIRM measurements, zero or low χ_{fd} values, and significantly lower ('harder') S-ratios. Although several 'spikes' were recorded in the χ_{fd} data, most of these are found to coincide with very low levels of χ_{lf} and χ_{hf} and, as discussed in section 6.28, are most likely artefacts of low susceptibility measurements. Figure 6.34 shows the variation of χ_{fd} as a function of χ_{hf}, demonstrating how the elevated χ_{fd} values in the lower part of the core correspond with, and are a consequence of, low susceptibility values. These sediments are therefore likely to be composed of lesser concentrations and/or volumes of ferrimagnetic minerals or antiferromagnetic minerals such as haematite, suggestive of lower deposition rates, more placid catchment conditions, and low rates of erosion. Low rates of deposition may have encouraged magnetic dissolution to take place, which would account for the absence of ultra-fine ferrimagnetic minerals. Another complementary explanation is that soil material transported to the basin in the past may have been magnetically distinct from that which dominated sediment yield at Tocal following European settlement. Some support for this theory comes from the study of the preserved pre-European soil from the western edge of the catchment (sections 6.8 and 7.8), which is both chemically and physically distinct from those soils found in the catchment today (section 7.8.5).

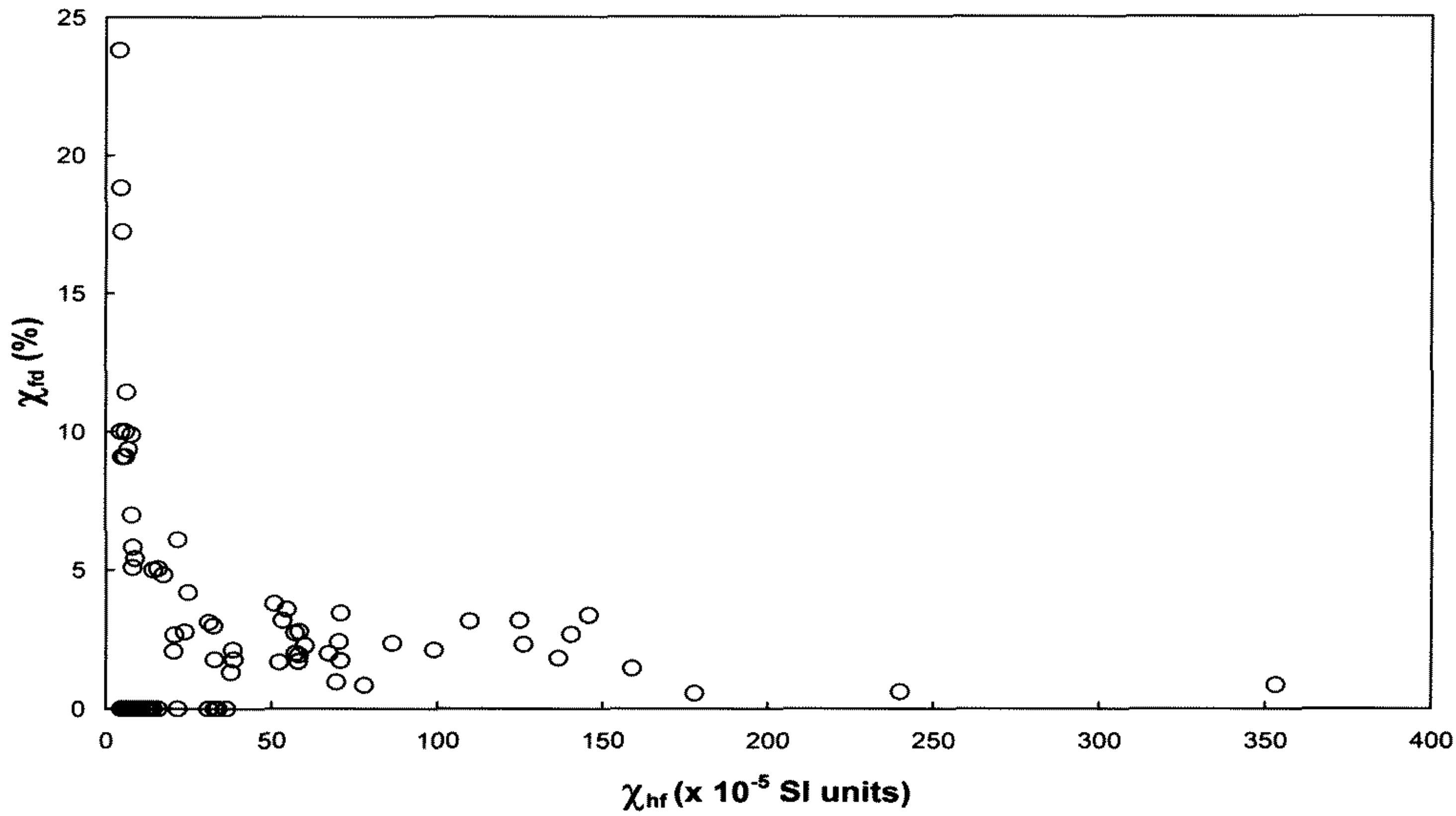


Figure 6.34 Variation of χ_{fd} as a function of χ_{hf}, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

6.11.2 Chemical indicators of environmental change

6.11.2.1 The interpretation of chemical profiles in the Tocal sediments

Variations in the ratios of P, Ca, Na, Mg, K and S are thought to provide the best potential sources of palaeoenvironmental information on environmental changes at Tocal. However, before changes in the elements can be interpreted as records of environmental change, the possibility of internal controls on these patterns must be investigated.

One possibility is that these patterns may be an artefact of changes in other elemental concentrations, due to the closed nature of geochemical data (Aitchison, 1986). However, the expression of each of these elements as ratios with the main elemental constituent of the sediment (SiO_2) shows that their concentrations fluctuate independently of this major element, and that the downcore pattern of the ratio is similar to that exhibited by the element alone (Figures 6.20 and 6.21).

Another possibility is that increases in these elements are related to lithological changes in the sediments. Little textural variation occurs in the Tocal sediments (section 6.2.1), particularly in the pre-European unit below 1.26 m in core TCA9b. Variations in these elements, such as the rise in phosphorus at 1.54 m depth, are therefore unlikely to be the result of textural changes.

Post-depositional processes may influence elemental records in lake sediments. Interaction between P and Fe, and to some extent Ca, has resulted in lake redox conditions influencing the sedimentary P record (Engstrom and Wright, 1984; Anderson and Rippey, 1994). Redox effects may be identified then by the coupling of P and Fe patterns in sediments (Engstrom and Wright, 1984). The downcore pattern of the P/Fe ratio in the Tocal sediments, however, shows no evidence of co-dependence (Figure 6.21). Perhaps the most widely invoked post-depositional process involving these elements is that of mobility within sedimentary sequences. Phosphorus, along with elements such as Ca, Na and K are potentially mobile in sediments. However, as Na (for example) exhibits a long-term decreasing trend in the sediments, while P displays a long-term increase, elemental mobility is unlikely in this system, as it would imply the movement of different elements in different directions in the sequence (Gale and Haworth, 2002). Thus, post-depositional mobility alone cannot account for the pattern of these elements in the Tocal sequence.

The possibility that increases in these elements in the sediments of Tocal Homestead Lagoon have been controlled by these internal processes may therefore be discounted. The patterns of elemental concentrations from core TCA9b, therefore, most probably record the history of their input into the lake basin from the surrounding catchment.

6.11.2.2 Phosphorus

Engstrom and Wright (1984), and many workers since, have detailed the use of P records in lakes to reconstruct past environmental conditions, and the problems involved in doing so. Under favourable conditions, P has proven useful as an indicator of past lake productivity (Olsson *et al.*, 1997; Boyle, 2001), as a record of nutrient input from the catchment (e.g., Schindler, 1974) and as a source of chronological information (e.g., Gale *et al.*, 1995). As nutrient levels in the Australian natural environment are typically low, P concentrations above background levels have proven to be an important marker of human and animal activity (Gale and Haworth, 2002; Gale *et al.*, 2004; Gale and Carden, 2005). Elevated P in sediments due to human and animal faeces, fertilisers and detergents has therefore been used as index of human and animal activity in the environment (Gale and Carden, 2005).

Major fluctuations in the P content of sediments from Tocal have been recorded during the last *c.* 2000 years. P concentrations first rise above background levels at 1.54 m (Figure 6.21). The only two environmental pathways for such an increase are enhanced detrital P input into the system during periods of increased catchment weathering or erosion, or an increase in human and/or animal activity or population in the catchment. The latter connection is based on the increase in organic P to the environment from human and animal faeces (Engstrom and Wright, 1984; Gale and Hoare, 1991). To complicate matters, there is good reason to believe that rises in P in the Tocal record may, at times, be driven by both processes in tandem. Increases in human activity, and thus organic P input into the system, often coincide with increased catchment disturbance and further increases in P in conjunction with parallel increases in cations such as calcium (Ca), potassium (K), magnesium (Mg) and sodium (Na) (sections 6.11.2.2 and 6.11.2.3). As the ratio of these cations to one another in clay minerals will be fixed, increases in erosion to the lake basin should be recorded as near equal rises in each element. Therefore, rises in P in the Tocal record which occur independent of these cations may record increases in human activity that are independent of catchment disturbance. The examination of how P varies against these cations may therefore allow a record of human activity to be distinguished from one of 'natural' catchment erosion (Figure 6.35).

As an example, the first significant rise in phosphorus above background levels in core TCA9b occurs at 1.54 m (Figure 6.35). The plot of the ratio of P/(Ca+K+Na+Mg) shows the same feature, suggesting that the rise of P is independent of the summed concentrations (and behaviour) of Ca, K, Mg and Na. This P increase then cannot be attributed to increases in sediment flux (which remain low through this period [section 8.2]) or catchment weathering. The only remaining explanation for this first rise in phosphorus above background levels must then relate to human activity in the catchment. Therefore, the study of both the total P (P/Si) and the non-detrital P (P/Fe, P/[Ca+K+Na+Mg]) content of the lake sediments may elucidate the relative contributions of phosphorus in the lake due to periods of erosion and human activity.

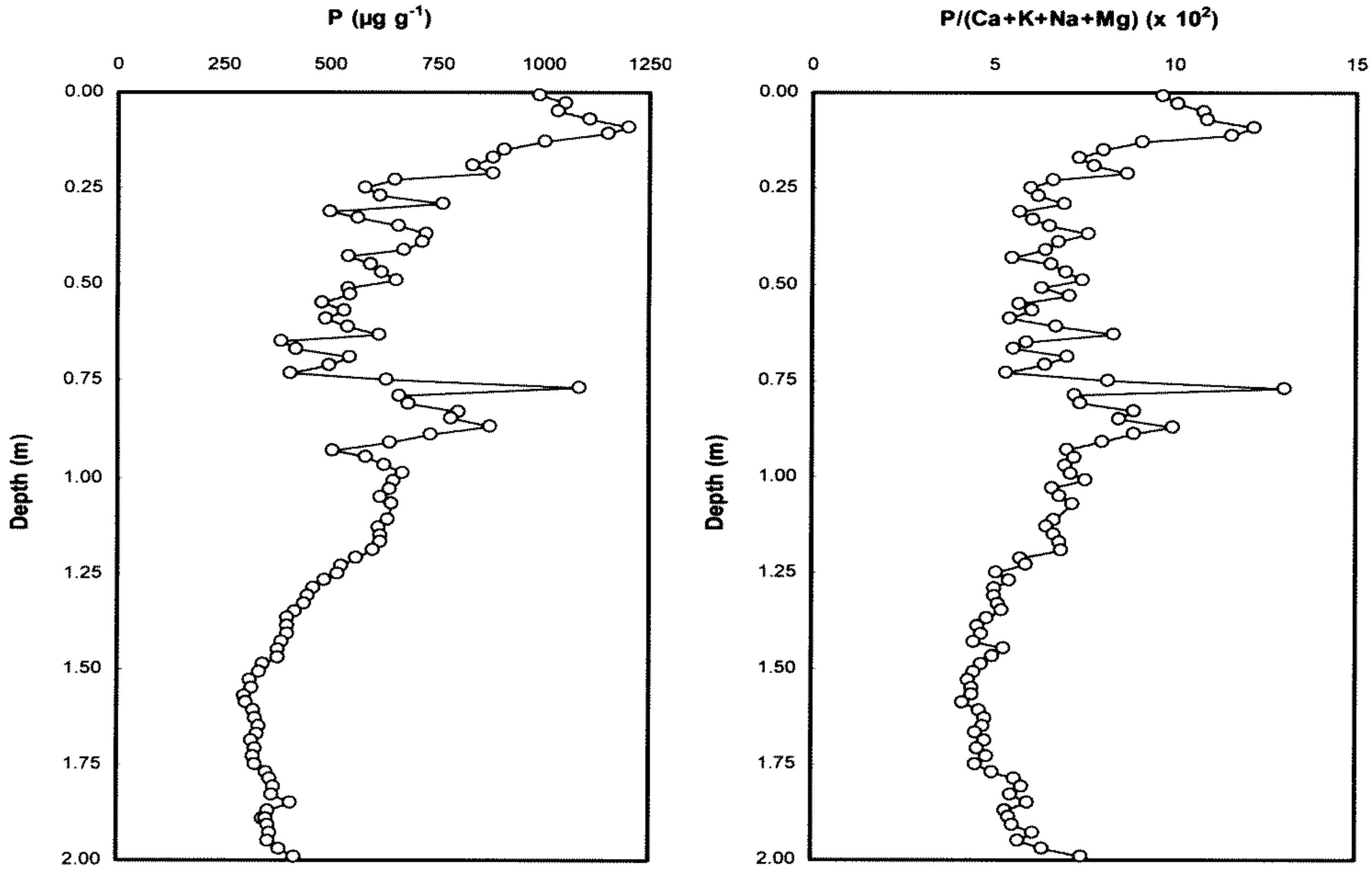


Figure 6.35 Downcore variation in P and the ratio of P with the sum of Ca, K, Na and Mg, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

6.11.2.3 Potassium and sodium

The cations potassium (K) and sodium (Na) are readily leached from alumino-silicate minerals in soils. After transport in solution, they are normally sorbed onto fine lake sediment particles (Engstrom and Wright, 1984). Records of K and Na in lake sediments due to catchment weathering may only become apparent in long-term records (Engstrom and Wright, 1984), which may be of little importance in the relatively short environmental record from Tocal. Under highly saline conditions K and Na may be precipitated at the mud-water interface (e.g., Valero-Garcés *et al.*, 1997), although this is highly unlikely to have occurred at Tocal. Many studies have related the accumulations of these elements in lakes to past erosion and weathering, and by extension to past climatic conditions (see Mackereth [1966], Engstrom and Wright [1984] and references therein). Pennington *et al.* (1972) provided evidence linking increased concentrations of K and Na in lake sediments to periods of increased catchment erosion. In the Tocal record, variations in K and Na (represented as their ratios with silicon [Figure 6.17]) will most probably record periods of past catchment stability and instability.

6.11.2.4 Calcium and magnesium

Calcium (Ca) and magnesium (Mg) are major components of the bulk chemistry of soils and bedrock. This is especially so in limestone terrain and coastal environments. Beyond these environments, the major source of Ca and Mg in lake sediments is similar to that of K and Na, clay minerals. As detrital input into lake basins, their behaviour is also similar to that of K and Na, increasing in concentration during periods of much greater erosion (Mackereth, 1966; Engstrom and Wright, 1984). In Mackereth's (1966) study from the Lake District of northwest England, Ca was found to be leached more easily from catchment soils than other alkali metals (K, Mg and Na), and peaks in sedimentary Ca were therefore only experienced during periods of greatest erosion, when the rate of erosion from the catchment was high enough to prevent the loss of Ca from the soils by leaching. Numerous studies worldwide have gone on to infer periods of catchment stability and instability based on patterns of Ca and Mg (and K and Na). Goundwater may also contribute Ca cations to lake sediments (Gale and Hoare, 1995). However, this alone cannot explain the downcore pattern in this element. Autochthonous sources of Ca and Mg are of importance to studies of long term lake palaeosalinity (e.g., Chivas *et al.*, 1986). However, authigneic Ca and Mg in lake sediments is related to shell material (Gale and Hoare, 1995; Cohen, 2003) and are of no importance at this site. Along with K and Na, the atmospheric input of Ca and Mg via dust deposition is unlikely, as rates of atmospheric dust deposition in the area are negligible (section 9.2.1). Thus, at Tocal, variations in these elements will be most closely linked to the erosion of minerogenic soil material into the lake.

6.11.2.5 Sulphur

Sulphur (S) may enter a lake basin via atmospheric, stream flow or groundwater sources, primarily as inorganic sulphates and S-rich organic compounds (Cohen, 2003). Another major source is the weathering or erosion of S-rich rocks and soil in catchments (Holmer and Storkholm, 2001). When sediments are derived from low-S bearing rocks and soils, and atmospheric deposition of S is not a major pathway, the greatest source of S input to lake sediments is from organic substances. In the presence of sufficient organic matter, S may be reduced to H_2S by bacteria on the lake bottom. Hydrogen sulphide may then react with sedimentary Fe to form various Fe-sulphide compounds or combine with organic matter which is subsequently buried (Cohen, 2003). Iron sulphides are quite insoluble under anoxic conditions (Cohen, 2003), and can accumulate in sedimentary sequences thus recording periods of higher organic productivity.

The formation of insoluble Fe sulphides in the sediment sequence reduces the binding of P–Fe oxides, and P may be released from the upper sediments into the lake, enhancing the eutrophic status of the lake (Holmer and Storkholm, 2001). In Lake Bussjösjön, southern Sweden, Olsson *et al.* (1997) reported the release of P from Fe-phosphates accompanying an increase in atmospheric S deposition in the lake. The remobilisation of P elevated the lake's productivity to its present highly eutrophic state. This study demonstrates the complex interactions that may be involved in changing elemental concentrations in lakes, and the role of different elemental phases in lake sediment geochemistry.

A study of a range of lakes by Urban *et al.* (1999) showed that the extent of S enrichment at the top of lake sediment sequences is correlated with lake trophic status. This is probably related to the organic matter required for initial sulphur reduction to take place. In any case, downcore changes in S concentrations, regardless of their specific origin, may record past changes in lake productivity. In such cases, the parallel examination of the organic matter content of sediments is needed to establish the connection, and to rule out diagenetic controls on S accumulation (Cohen, 2003). Also, the contribution of atmospheric S deposition from industrial pollutants must be known at a site and whether possible shifts in sediment provenance to S-richer material may have taken place.

At Tocal, the concentration of S in the lake's sediments displays distinct peaks and troughs (Figure 6.36). No major sources of sulphur exist in the Tocal catchment, and peaks of S in the lake sediments cannot be due to either increases in the erosion of S-rich soils. An atmospheric pollution source for this pattern is also unlikely. There are no sources of industrial S pollution in the region, and even if there were, none would be capable of explaining the approximately 100% increase which dates from the early 19th century AD (above 1.24 m) (Figure 6.36).

The peaks of S in the lake sediments, therefore, are probably organic in origin. During periods of elevated organic productivity in the lake, organic-rich material and Fe sulphides formed and were subsequently buried. A way in which this thesis may be tested is by comparison with the LOI_{430} estimates of plant organic matter in the lake (Figure 6.36). Each major peak in sulphur coincides with peaks in LOI_{430}, showing a strong connection with former periods of enhanced lake productivity. The downcore trend in S also mirrors that of P (Figure 6.35), suggesting a relationship with nutrient input into the lake.

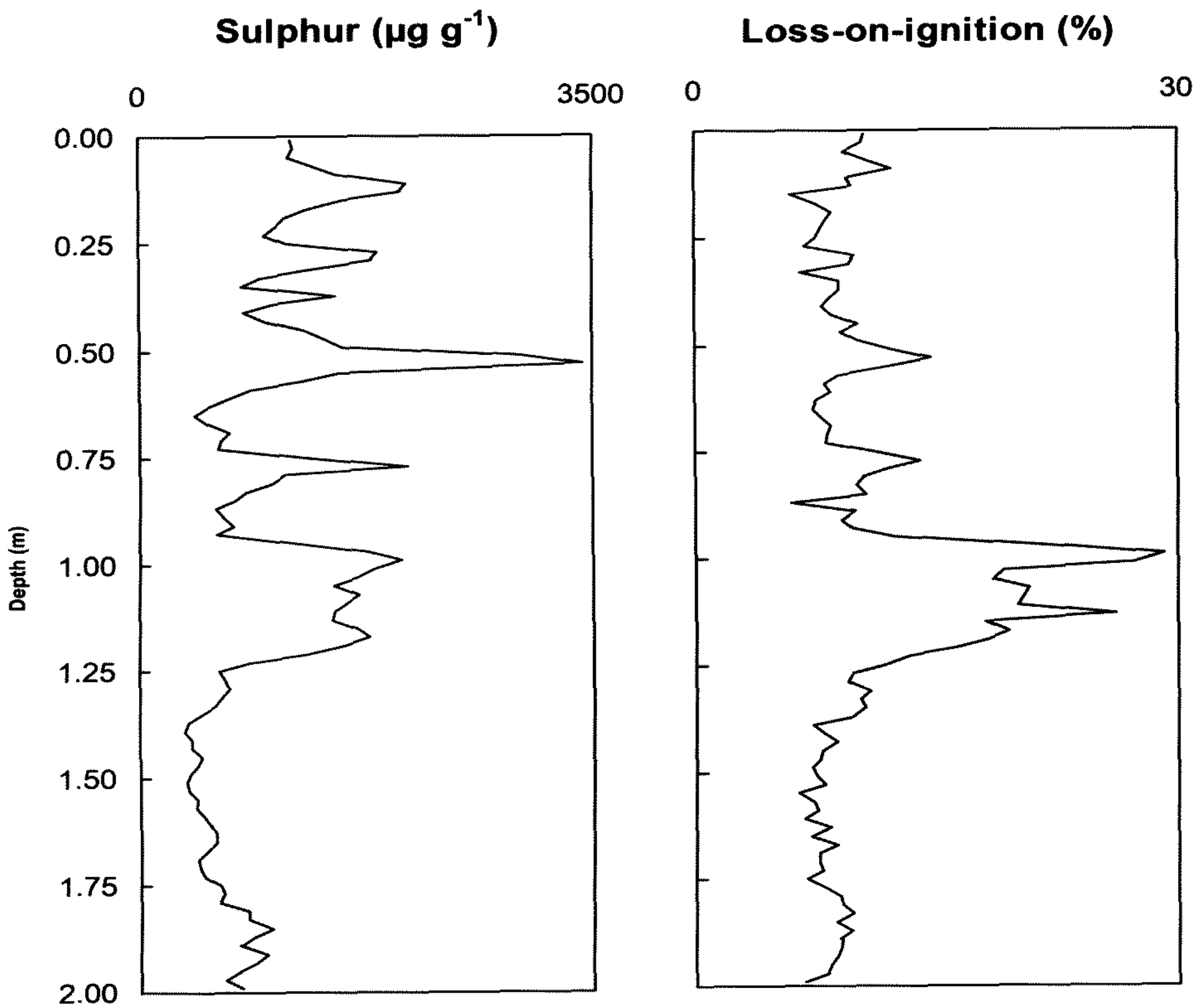


Figure 6.36 Downcore variation in S and loss-on-ignition, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

As the Tocal S record mirrors the increase in LOI_{430} content of the sediments, and is associated with increasing P content, the origin of this S is almost certainly authigenic S-rich organic compounds (Urban *et al.*, 1999; Cohen, 2003). The S record at Tocal may therefore be cautiously interpreted as a record of changing trophic status of the lake (Cohen, 2003). Increasing diagenetic S deposition during periods of high productivity may have further exacerbated the lake's eutrophic state, as increasing S loads in lakes may have triggered the release of additional P from the lake sediments.

6.11.3 Loss-on-ignition

Measurements of LOI_{430} (section 5.3.4) provide an estimate of the plant organic content (mainly organic carbon) of a material (Gale and Hoare, 1991). The loss in mass of a sample using this procedure is thought to be greater than the true amount of plant matter, due to the loss of hydroxyl ions and water of hydration (Smith and Atkinson, 1975; Gale and Hoare, 1991). Materials with a significant carbonate component may also suffer a loss of carbon dioxide (Gale and Hoare, 1991) and LOI_{430} measurements made on carbonate-rich material will therefore provide a poor estimate of plant organic matter. The loss of structurally bound water during LOI_{430} measurements is of particular concern in clay samples, where 10% by mass of the sample may be structural water (Boyle, 2001). Gale and Carden (2005), for example, have suggested that very low LOI_{430} values (1.4–1.7%) in sandy sediments reflected the loss of hydroxyl ions and water of hydration rather than the ignition of plant organic matter. Thompson and Eglinton (1978), and others, have shown that organic matter content is generally low in coarser sedimentary material and increases with decreasing sediment grain size.

In Tocal Homestead Lagoon, a comparison of the downcore variation of LOI_{430} (Figure 6.5) section with the visual logs of the sediments (Figure 6.3 and 6.4) suggests that plant matter is predominantly composed of very fine plant stems, roots and rootlets in growth position. Although it is inevitable that some allogenic organic material has been eroded into the basin in the past, other than charcoal none has been identified visually, and it is considered to be negligible in comparison with these sources. A possible particle-size control on the downcore pattern of LOI_{430} is also unlikely given the lack of textural changes in the sequence. Therefore, changes in LOI_{430} values almost certainly represent the amount of aquatic plant material which grew and was buried on the lake bed.

Chapter Seven

Chronology

7.1 Introduction

This chapter draws upon several of the results from the laboratory analysis of the mastercore TCA9b to provide information on the age of the sediments in Tocal Homestead Lagoon. In particular, the results of the analysis of these sediments for their radiometric properties (^{14}C, ^{137}Cs and ^{210}Pb), elemental lead content and palaeomagnetism will be discussed and interpreted for chronological information. Those sediments laid down prior to the arrival of Europeans at Tocal in the early 19th century are dated using ^{14}C methods. Although a very common means of dating sediments of up to, and beyond, 40 000 years of age, this procedure is not without complexities. Some complications involved with ^{14}C dating sediments at Tocal have been overcome by the parallel dating of different materials from the sediments.

The main age information for this sequence is provided by the modelling of excess ^{210}Pb activities, which can provide age estimates for sediments deposited within the last 200 years (e.g., Gale *et al.*, 2004). However, the use of such data to provide sediment ages is not straightforward, and a range of models is available which can generate downcore profiles of sediment ages. To ensure that it provides realistic estimates of sediment ages, any model used should be tested against some independent time-marker in the sediments (Smith, 2001). In this research, I have undertaken three separate experiments to obtain independent information on the chronology of sedimentation over the last two centuries. These have involved the use of ^{137}Cs activities, elemental lead records and sediment palaeomagnetism to test the sediment ages estimated from the modelling of the downcore flux–activity of excess ^{210}Pb.

7.2 Caesium-137 chronology

7.2.1 Caesium-137 as a time-marker

Caesium-137 is an artificial radioactive isotope produced and dispersed as a by-product of nuclear weapons testing. Most of this testing was undertaken in the northern hemisphere, beginning on a significant scale after World War 2. Since then, atmospheric nuclear weapons testing has released the isotope into the stratosphere, from where it has been redistributed over the Earth's surface as fallout. The peak of ^{137}Cs supply to the northern hemisphere is known to be sharply focused in 1962–1965 (Figure 7.1). Observations of the ^{137}Cs activity of lake sediment sequences in the northern hemisphere have demonstrated their ability to record the historical flux of ^{137}Cs, with the first appearance and peak activity of ^{137}Cs dated to 1954 and 1963 respectively (e.g., Pennington *et al.*, 1973; Robbins and Edgington, 1975).

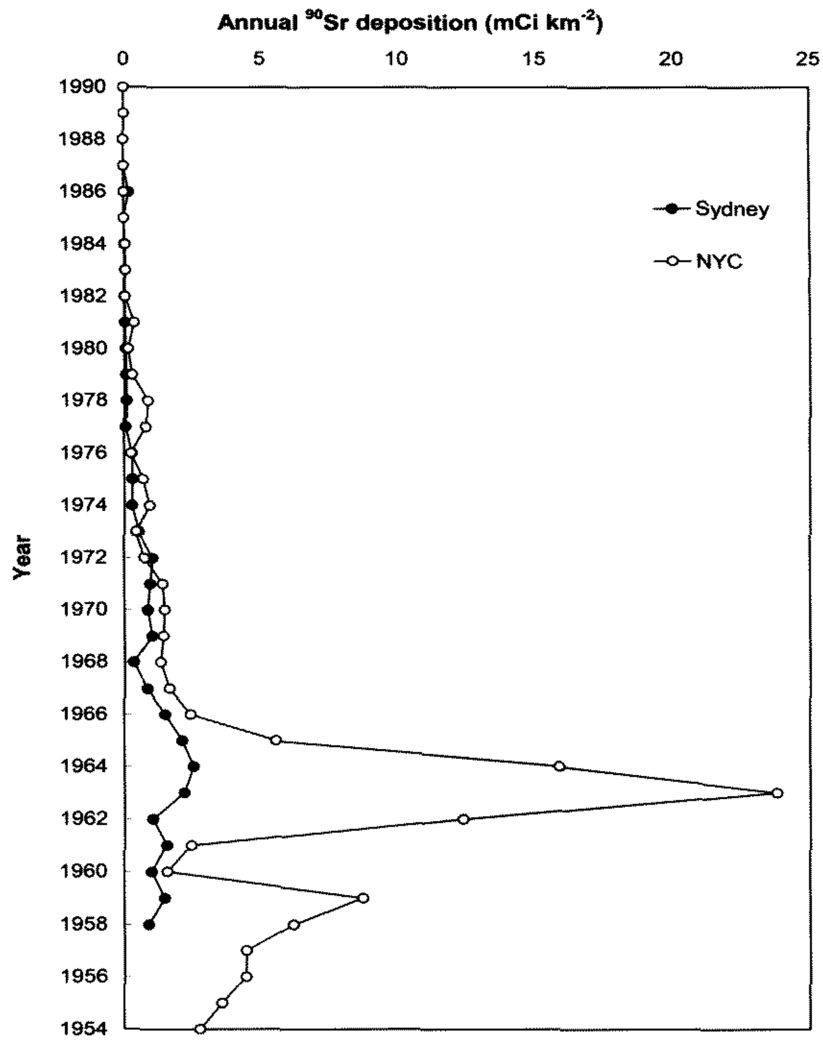


Figure 7.1 Annual ^{90}Sr atmospheric deposition in Sydney, Australia and New York City, USA for the period 1954 to 1990. Data for Sydney were not collected until 1958. No ^{137}Cs fallout data are available for Sydney, so the pattern of fallout of ^{90}Sr for the two cities is used to show the differences in deposition in the northern and southern hemispheres. The data are from the Environmental Measurements Laboratory website (http:www.eml.doe.gov/databases), accessed 3/3/04. Based on a similar chart produced by Brunskill and Pfitzner (2002) for Brisbane, Australia.

The history and magnitude of radionuclide fallout in Australia is quite different to that of the northern hemisphere, with the peak input of ^{137}Cs only two times greater than that for the 1950–1960 or 1980–present periods (Figure 7.1) The peak flux of ^{137}Cs in Australia was maintained over a much greater period than in the northern hemisphere, 1954–1974, due to frequent, but much smaller nuclear bomb detonations in the 1950s and early 1970s (Murorora, Maralinga, Monte Bello Island) (Brunskill and Pfitzner, 2002). As a result, undisturbed records of ^{137}Cs in Australia would not be expected to display the sharp 1963 peak, but should instead exhibit a relatively flat curve with minor peaks (Longmore, 1982). A consensus in most published Australian studies is that the peak in ^{137}Cs activity recorded in the northern hemisphere cannot be used as a datable horizon in Australian soils and sediment and that only the first appearance of ^{137}Cs may be identified with any confidence (Olley *et al.*, 1991). Based on the performance of the gamma-ray detectors and the counting system used at CSIRO, Canberra, and taking into account the characteristics of these materials that were typically analysed, Brunskill *et al.* (2002a, p. 250), calculated the date of the first detectable appearance of ^{137}Cs in Australian fine-grained soil and sediment as 1958. Importantly, Brunskill *et al.* (2002a) noted that, because the input of ^{137}Cs into the Australian environment has effectively ceased, the activity of the ^{137}Cs remaining in the landscape is decreasing over time. This means that the date allocated to the depth of first detection is not fixed, but is increasing in age at a rate controlled by the isotope's half-life and its past history of input into the Australian environment.

The dating capabilities of the ^{137}Cs method rely on the assumption that ^{137}Cs is immobile, and therefore remains in a stratigraphic sequence at the point of its deposition. Numerous studies have shown that ^{137}Cs is rapidly and firmly attached to soil and sediments (Longmore [1982] and references therein). Nevertheless, caesium has been shown to be mobile in sediments of soft-water lakes with relatively high organic matter (>20% LOI_{430}), and sediments of low clay mineral content (Oldfield *et al.*, 1979; Davis *et al.* 1984; Davison *et al.*, 1993). Other problems that have been experienced with the technique include the lag-time that exists between the direct atmospheric deposition of ^{137}Cs and the erosion of ^{137}Cs bound to soil material into a lake system (Ritchie *et al.*, 1973). This process gives rise to a vertical smearing of the ^{137}Cs in a sedimentary sequence, hampering attempts to use it as a time-marker. Ritchie *et al.* (1973) demonstrated this lag to be between 6 and 12 months in small catchments, an error that would be detectable only in extremely high-resolution studies of rapidly sedimenting systems. Other problems such as those associated with reworking by wave action or bioturbation are unrelated to the dating technique, and are best solved by careful site selection and sample collection programmes.

7.2.2 Caesium-137 dating in Australia

The use of ^{137}Cs as a dating technique in Australian research has been relatively uncommon. Perhaps this has been in part due to poor results obtained in dating fluvial sediments (e.g., Bishop *et al.*, 1991) and lake sediments unsuited to the technique (e.g., Torgersen and Longmore, 1984). Caesium-137 has been used successfully to date sediments in a farm dam in the Hunter valley of eastern New South Wales (Loughran and Campbell, 1983), Burrinjuck Reservoir in southeast New South Wales (Wasson *et al.*, 1987), Lake Albert, southern New South Wales (Caitcheon, 1990), Lake Illawarra, eastern New South Wales (Chenhall *et al.*, 1995), Redhead Lagoon, eastern New South Wales (Williams, 2005), and Lake Alexandria, South Australia (Herczeg *et al.*, 2001).

7.2.3 The ^{137}Cs chronology of core TCA9b

Some information on the age of sediments from the upper part of the Tocal lake sediment sequence comes from measurements of ^{137}Cs activities in core TCA9b. In Australia, it is usually assumed that the occurrence of ^{137}Cs at detectable levels took place only after the late-1950s (Brunskill *et al.*, 2002a). The ^{137}Cs determinations made on the core are reproduced in Figure 7.2. No ^{137}Cs was detected below 0.44 m in the sequence. Therefore, the sediments from this depth could have been deposited no earlier than 1958 and most probably date from this time

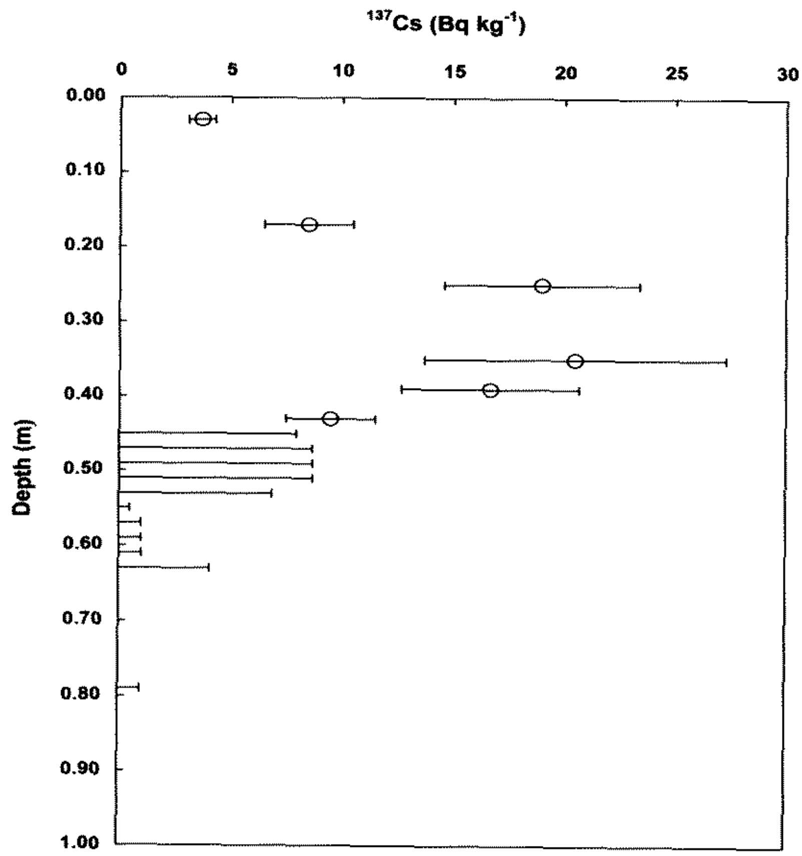


Figure 7.2 Downcore variation in ^{137}Cs activity, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Vertical error bars span sample slice depths (0.02 m), while horizontal error bars represent 66% confidence limits for each measurement. The replicate measurement of sample 0.56–0.54 m confirmed no detectable ^{137}Cs activity at this depth.

7.3 Lead pollution as an independent time-marker

(from Cook and Gale, 2005)

7.3.1 Lead pollution histories

Although heavy-metal pollution histories are well defined at many sites in the northern hemisphere (e.g., Nriagu, 1983; Shotyk *et al.*, 1998, Monna *et al.*, 2004), and at a limited number of locations in the southern hemisphere (e.g., Antarctica [Planchon *et al.*, 2003]) there exists a paucity of such histories for Australia. Those few studies that exist have often been based on a records of atmospheric deposition of metals in soils and sediments (e.g., Gale *et al.*, 1995; Chiaradia *et al.*, 1997), utilising those few historical records that exist to corroborate the preserved evidence.

Of particular interest in this context have been those attempts to unravel the history of lead input to the environment through time. The role of lead pollution in environmental contamination has been the subject of immense scrutiny by researchers, governments and community groups alike, due to its toxicity to humans particularly to infants (Nriagu, 1978). Lead is relatively immobile in soils and sediments, and is persistent in the environment over very long time-scales, meaning that past accumulations of toxic lead may be reworked to become a modern day problem.

In Australia, the first inputs of anthropogenic lead into the atmosphere and terrestrial environments following European settlement in the late 18th century probably consisted of dusts ejected into the atmosphere as the result of coal and lead mining, while lead smelting caused significant localised lead pollution by the late 19th century. However, the greatest single event in the history of lead pollution was undeniably the introduction of lead additives into fuels. By the end of the 20th century, Nriagu (1990) estimated that the lead input into the atmosphere by emissions of motor vehicles constituted the greatest source of lead atmospheric pollution worldwide, accounting for 61% of the total lead atmospheric pollution in the late 1980s.

7.3.2 The first use of leaded fuel in Australia, 1932 AD

Although it is well known that the first leaded fuel left the pump in February 1923 in the United States (Nriagu, 1990, p. 15), the chronology of its introduction elsewhere is much less well understood. Australia is of some significance here as it possessed over a third of the motor vehicles in the southern hemisphere until well into the second half of the 20th century (Figure 7.3). It is thus likely to have been the major single source of atmospheric tetraethyl lead in the southern hemisphere during that period. Unfortunately, there is little agreement over when leaded fuel made its appearance in Australia.

Winser (*c*. 1955, p. 37), for example, noted that the introduction of 'Ethyl superpetrols' took place in 1928, although he later wrote that the first leaded fuel introduced to Australia was 'Super Plume' in 1934 (Winser, *c*. 1955, p. 262). In his history of Australian motoring, Davis (1987, pp. 63–64) reported that the Commonwealth Oil Refineries pioneered the production of leaded fuels in Australia in 1924. By contrast, Arnold's (2000, p. 26) history of Australian petrol companies contended that leaded fuel was introduced in 1934, initially by the Commonwealth Oil Refineries and later that same year by the Atlantic Union Oil Company (Arnold, 2000, p. 22) and the Vacuum Oil Company (Arnold, 2000, p. 92). In their study of lead pollution around Lake Illawarra in eastern New South Wales, Chiaradia *et al.* (1997, p. 112) noted that lead was first added to fuel in 1948. Finally, the Australian Institute of Petroleum reported that leaded fuel was introduced to Australia in the 1920s (Ewen Macpherson, Deputy Director, Australian Institute of Petroleum, written communication, 4 June 2002). Unfortunately, none of these contradictory and apparently authoritative sources, whose assessments cover a range of over 20 years, are correct.

The Anglo-American Oil Company introduced leaded motor fuel to the British retail market in January 1928 (Bamberg, 1994, pp. 195–196). However, the availability of relatively high-quality blended fuels in Britain, particularly by comparison with the USA, reduced the incentive to develop leaded petrol (Anon., 1930), and it was not until 15 April 1931 that British Petroleum (BP) began marketing leaded fuel (Bamberg, 1994, p. 196). The fuel, known as '"B.P." Plus', was tinted blue to distinguish it from unleaded petrol (Whatmough, 1931, p. 229). The new fuel was an immediate success and BP's sales increased markedly (Anon., 1932a). Following this lead, Australia's Commonwealth Oil Refineries (COR), 49% of which was owned by BP (Ferrier, 1982, p. 522), introduced the first leaded fuel to Australian consumers in 1932. The new petrol was marketed as '"C.O.R." Plus' and was probably similar, if not identical, to the '"B.P." Plus' sold in Britain.

The history of the introduction of leaded fuel to Australia is recorded in *The Accelerator*, the house journal of COR. In late July 1932, a representative of the Ethyl Export Corporation arrived in Australia to supervise the initial blending of '"C.O.R." Plus' (Anon., 1932b). Nevertheless, it was not until the publication of the October 1932 issue of *The Australian Motorist* that it was formally announced that 'A new blue petrol is now being distributed by the Commonwealth Oil Refineries Ltd.' and that 'The blue colour ... denotes the presence of a special lead compound ...' (Anon., 1932a). The new fuel is likely to have been on the market some time prior to this date, however, for the first advertisement for '"C.O.R." Plus' had appeared on p. 43 of the September 1932 issue of *The Australian Motorist* (Figure 7.3). Moreover, reports from COR customers in the September and October issues of *The Accelerator* suggest that the fuel had been available to the public at least as early as August (Anon., 1932c; 1932d). The success of the new fuel saw the rapid introduction of other brands of leaded petrol to the Australian market, although even as late as April 1933 '"C.O.R." Plus' was still the only leaded fuel commercially available in Australia (George, 1933). The first advertisement for the Atlantic Union Oil Company's 'Atlantic Ethyl' appeared in the September 1934 (p. 29) issue of the journal *The Motor in Australia and Flying*. A month later, advertisements for the Vacuum Oil Company's 'Super Plume Ethyl' appeared in *The Motor in Australia and Flying* (p. 28) and *The Australian Motorist* (p. 101). These were followed on 1 April 1935 by a full-page advertisement in *The Australian Motorist* for the Ethyl Export Corporation (p. 416), the company that produced the tetraethyl lead used in leaded fuels. This pointed out that 'Ethyl petrol is now available to all Australian motorists from pumps bearing the Ethyl trade mark'.

Leaded petrol marketed by the Commonwealth Oil Refineries was thus being sold in Australia by August 1932. The speed with which other companies introduced similar brands of fuel testifies to the rapid take-up of leaded petrol by Australian motorists. Although the immediate environmental impact of leaded fuel is likely to have been highly localised, Australians took to cars early and in relatively large numbers. In New South Wales alone, for example, there were 212 137 vehicles registered in December 1932 (Anon., 1933). Correspondence in *The Accelerator* of 1932–1933 reported long distance and speed trials performed by motorists across the country using COR's new fuel, and leaded fuel soon became available in parts of the country remote from the capital cities. By *c.* 1938, for example, 'Super Plume Ethyl' was on sale in Deniliquin, 600 km inland of Sydney (Anon., *c.* 1938) (Figure 7.5).

Figure 7.3 Announcement of the introduction of leaded motor fuel to the Australian market. The advertisement appeared on p. 43 of the 1 September 1932 issue of *The Australian Motorist*.

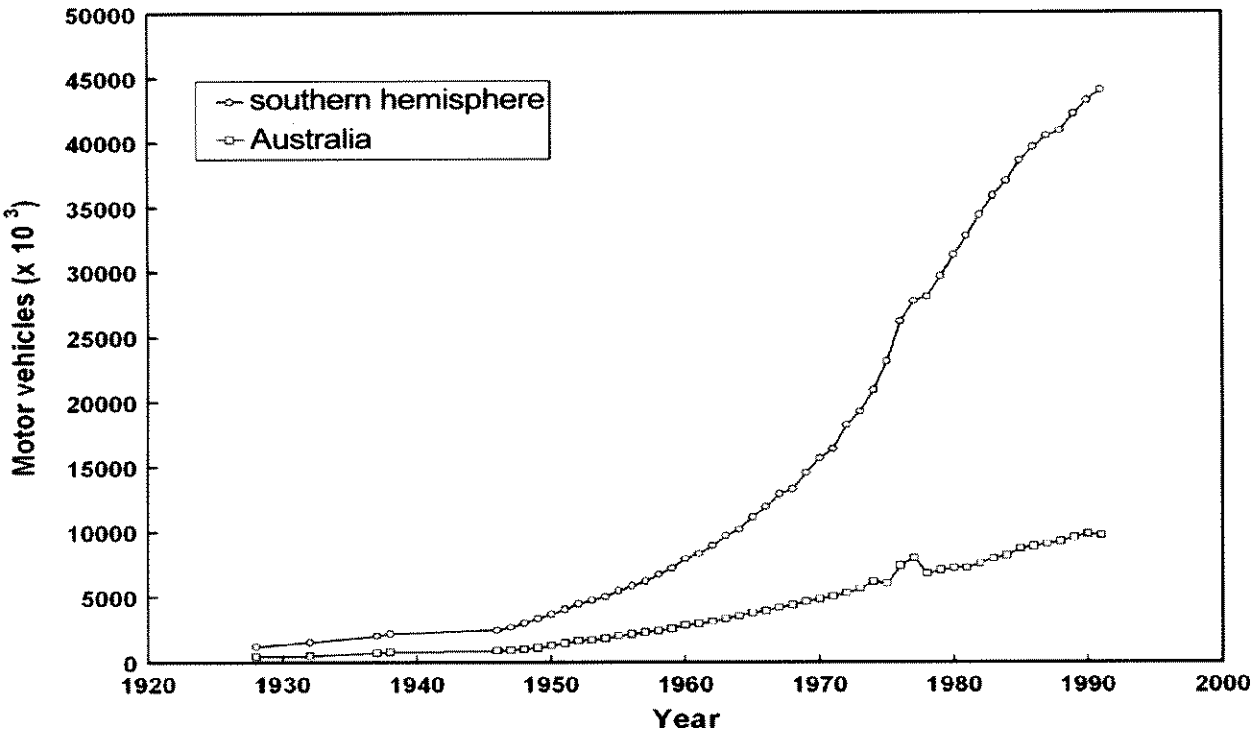


Figure 7.4 Number of motor vehicles in Australia and the southern hemisphere, 1928–1991. Source of data: Statistical Office of the United Nations and its successors (1949–1955, 1957, 1959–1961, 1963–1979, 1981, 1983, 1985, 1988, 1992–1997, 1999–2002). The data were obtained by summing individual returns for all southern hemisphere countries. Occasional gaps in the data set have been filled by interpolation and extrapolation. From Cook and Gale (2005).

The relatively high fuel consumption and substantial exhaust emissions characteristic of vehicles of the time would have meant that each car was capable of injecting large quantities of lead into the environment. Over much of the country, therefore, it is likely that the take-off in environmental lead levels dates from this time. The introduction of leaded fuel to Australia may also be significant on a hemispheric scale, since for most of the 20th century Australia possessed significantly more motor vehicles than any other southern hemisphere country (Figure 7.4). Australia alone is thus likely to have contributed around a third of the southern hemisphere's tetraethyl lead fallout.

Figure 7.5 'In End Street – Deniliquin, NSW', *c.* 1938. This image is from State Library of New South Wales Photograph Collection, At Work and Play – Images of Rural Life in NSW 1880–1940 Frame Number 06190, Sydney. The second petrol pump from the left is labelled 'Super Plume Ethyl', confirming the availability of leaded fuel in remote locations in Australia shortly after its launch.

7.3.3 The record of atmospheric lead at Tocal Homestead Lagoon

The concentration of lead recorded in the sediments at Tocal shows a clear pattern of increase above background levels above 1.00 m (Figure 7.6). The ratio of Pb with Si is also shown here, demonstrating that the Pb record is likely to provide a real measure of the lead content of the sediments, rather than being an artefact of the complexities of compositional data. Lead of detrital origin is characterised by a consistent ratio with the major components of the soil, such as silicon. The only possible explanation for an increase in lead above this relatively constant 'background' level, therefore, is anthropogenic; no shift in sediment source or limnological process could explain such a sharp rise in heavy metal levels in the undisturbed sedimentary sequence from Tocal Homestead Lagoon. As discussed in section 7.3.2, major increases in atmospheric lead levels in the environment in Australia began in 1932 with the introduction of leaded fuel. At Tocal, motor vehicles were in use in the catchment through the 1930s. These included a 1934 Rolls Royce owned by C.B. Alexander that would have used leaded fuel (section 4.7.2). The rise in sedimentary lead levels at Tocal is thus likely to reflect the dramatic increase in atmospheric lead fallout associated with the introduction of tetraethyl lead in fuels in 1932. The depth in the sequence at which the first sediments with elevated lead levels are preserved is not altogether clear in the downcore profile, and cannot be simply identified visually. Lead levels increase sharply, but continually above *c.* 1.00 m. Examination of these sediments as two populations, one representing pre-increase conditions, the other post-increase, is required. The highest sediment that is not a member of the 'background lead' population is most probably the first sediment to contain lead levels above the background.

7.3.4 Identifying the first increase in lead levels in core TCA9b

The mean and standard deviation of sequence of samples thought to contain only pre-tetraethyl lead sediments from core TCA9b was calculated (Table 7.1). Each group was checked to confirm whether the lead concentrations of all group members lay within two standard deviations of the group's mean. If this was so, then those sediments could be considered to have no lead concentrations above background. Subsequent tests would then include the next sample depth up the sequence. The lead concentrations in the sediments from 2.00–0.92 m all lie within two standard deviations of the mean of this 'background' group. The sediment from 0.92–0.90 m has a lead concentration of 16.0 $\mu g\ g^{-1}$, however, making it the stratigraphically lowest sample to be statistically distinguishable from the 'background' group (Table 7.1). However, this is based on the mean lead concentration for this depth. When the error range of lead in this sample (minimum concentration of 15.2 $\mu g\ g^{-1}$) is taken into account, the sample may be seen to lie within the range of background lead levels for the group 2.00–0.90 m (whose members have a maximum concentration of 15.9 µg g^{-1} at 95% confidence limits). Taking measurement errors into account, the stratigraphically lowest sediments possessing lead concentrations above the range of background values are those from 0.84–0.82 m (19.9 ± 1.0 4 µg g^{-1}) (Table 7.1 and Figure 7.6).

Utilising an entirely separate statistical approach, the within group and between group distances in the Pb data for the sequence were examined using three different clustering techniques (K-means, Ward's and single linkage methods). All three procedures separate data into two agglomerates; those of background lead concentration, and those with anthropogenically elevated lead levels. In every case the lowermost sample defined as being affected by lead pollution is 0.84–0.82 m (Table 7.2).

All four methods indicated that the depth at which the lead content first increases above background concentrations lies at 0.84 m; that is, the sample from 0.84–0.82 m is the first to contain lead levels statistically distinguishable from the natural background of the site. Since the greatest increase in atmospheric fallout of lead at the site almost certainly took place at or shortly after 1932, a date of *c.* 1932 may be assigned to 0.84 m in the core.

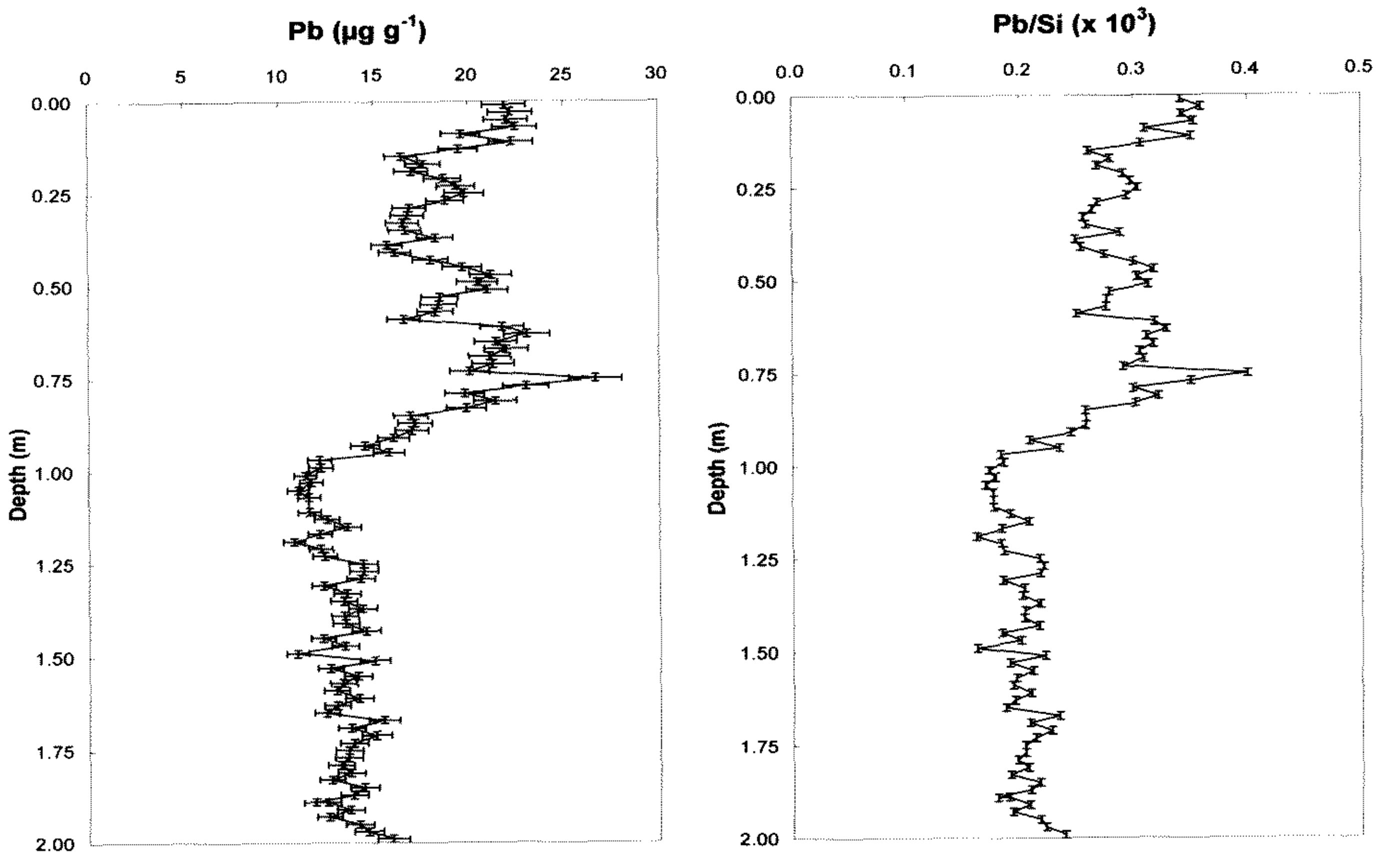


Figure 7.6 Downcore variation of Pb and the Pb/SiO_2 ratio, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. To better identify changes in the sequence, a measure of error on the lead determinations has been provided. Based on repeat measurements, the 95% probability range of lead concentrations for each sample depth is ± 5.22% (horizontal error bars). Vertical error bars represent sample depths (0.02 m).

Table 7.1 Examination of the 'background' lead concentrations of groups of sediment, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Ten scenarios are examined, in which the mean and 95% confidence levels for each background lead group are calculated.

	Lead concentrations (µg g^{-1}) in background lead groups									
Sediment depth (m)	2.00 – 1.00 m	2.00 – 0.98 m	2.00 – 0.96 m	2.00 – 0.94 m	2.00 – 0.92 m	2.00 – 0.90 m	2.00 – 0.88 m	2.00 – 0.86 m	2.00 – 0.84 m	2.00 – 0.82 m
0.84–0.82										19.89
0.86–0.84									16.93	16.93
0.88–0.86								17.15	17.15	17.15
0.90–0.88							16.98	16.98	16.98	16.98
0.92–0.90						16.01	16.01	16.01	16.01	16.01
0.94–0.92					14.51	14.51	14.51	14.51	14.51	14.51
0.96–0.94				15.78	15.78	15.78	15.78	15.78	15.78	15.78
0.98–0.96			12.12	12.12	12.12	12.12	12.12	12.12	12.12	12.12
1.00–0.98		12.20	12.20	12.20	12.20	12.20	12.20	12.20	12.20	12.20
1.02–1.00	11.38	11.38	11.38	11.38	11.38	11.38	11.38	11.38	11.38	11.38
1.04–1.02	11.70	11.70	11.70	11.70	11.70	11.70	11.70	11.70	11.70	11.70
1.06–1.04	11.01	11.01	11.01	11.01	11.01	11.01	11.01	11.01	11.01	11.01
1.08–1.06	11.58	11.58	11.58	11.58	11.58	11.58	11.58	11.58	11.58	11.58
1.12–1.10	11.56	11.56	11.56	11.56	11.56	11.56	11.56	11.56	11.56	11.56
1.14–1.12	12.49	12.49	12.49	12.49	12.49	12.49	12.49	12.49	12.49	12.49
1.16–1.14	13.59	13.59	13.59	13.59	13.59	13.59	13.59	13.59	13.59	13.59
1.18–1.16	12.12	12.12	12.12	12.12	12.12	12.12	12.12	12.12	12.12	12.12
1.20–1.18	10.78	10.78	10.78	10.78	10.78	10.78	10.78	10.78	10.78	10.78
1.22–1.20	12.17	12.17	12.17	12.17	12.17	12.17	12.17	12.17	12.17	12.17
1.24–1.22	12.35	12.35	12.35	12.35	12.35	12.35	12.35	12.35	12.35	12.35
1.26–1.24	14.39	14.39	14.39	14.39	14.39	14.39	14.39	14.39	14.39	14.39
1.28–1.26	14.42	14.42	14.42	14.42	14.42	14.42	14.42	14.42	14.42	14.42
1.30–1.28	14.25	14.25	14.25	14.25	14.25	14.25	14.25	14.25	14.25	14.25
1.32–1.30	12.32	12.32	12.32	12.32	12.32	12.32	12.32	12.32	12.32	12.32
1.34–1.32	13.55	13.55	13.55	13.55	13.55	13.55	13.55	13.55	13.55	13.55
1.36–1.34	13.39	13.39	13.39	13.39	13.39	13.39	13.39	13.39	13.39	13.39
1.38–1.36	14.37	14.37	14.37	14.37	14.37	14.37	14.37	14.37	14.37	14.37
1.40–1.38	13.41	13.41	13.41	13.41	13.41	13.41	13.41	13.41	13.41	13.41
1.42–1.40	13.46	13.46	13.46	13.46	13.46	13.46	13.46	13.46	13.46	13.46
1.44–1.42	14.53	14.53	14.53	14.53	14.53	14.53	14.53	14.53	14.53	14.53
1.46–1.44	12.27	12.27	12.27	12.27	12.27	12.27	12.27	12.27	12.27	12.27
1.48–1.46	13.42	13.42	13.42	13.42	13.42	13.42	13.42	13.42	13.42	13.42
1.50–1.48	10.94	10.94	10.94	10.94	10.94	10.94	10.94	10.94	10.94	10.94
1.52–1.50	15.00	15.00	15.00	15.00	15.00	15.00	15.00	15.00	15.00	15.00
1.54–1.52	12.68	12.68	12.68	12.68	12.68	12.68	12.68	12.68	12.68	12.68
1.56–1.54	14.12	14.12	14.12	14.12	14.12	14.12	14.12	14.12	14.12	14.12
1.58–1.56	13.34	13.34	13.34	13.34	13.34	13.34	13.34	13.34	13.34	13.34
1.60–1.58	12.95	12.95	12.95	12.95	12.95	12.95	12.95	12.95	12.95	12.95
1.62–1.60	14.15	14.15	14.15	14.15	14.15	14.15	14.15	14.15	14.15	14.15
1.64–1.62	13.00	13.00	13.00	13.00	13.00	13.00	13.00	13.00	13.00	13.00
1.66–1.64	12.46	12.46	12.46	12.46	12.46	12.46	12.46	12.46	12.46	12.46
1.68–1.66	15.48	15.48	15.48	15.48	15.48	15.48	15.48	15.48	15.48	15.48
1.70–1.68	13.76	13.76	13.76	13.76	13.76	13.76	13.76	13.76	13.76	13.76
1.72–1.70	15.08	15.08	15.08	15.08	15.08	15.08	15.08	15.08	15.08	15.08
1.74–1.72	13.91	13.91	13.91	13.91	13.91	13.91	13.91	13.91	13.91	13.91
1.76–1.74	13.65	13.65	13.65	13.65	13.65	13.65	13.65	13.65	13.65	13.65
1.78–1.76	13.63	13.63	13.63	13.63	13.63	13.63	13.63	13.63	13.63	13.63
1.80–1.78	13.19	13.19	13.19	13.19	13.19	13.19	13.19	13.19	13.19	13.19
1.82–1.80	13.73	13.73	13.73	13.73	13.73	13.73	13.73	13.73	13.73	13.73
1.84–1.82	12.73	12.73	12.73	12.73	12.73	12.73	12.73	12.73	12.73	12.73
1.86–1.84	14.44	14.44	14.44	14.44	14.44	14.44	14.44	14.44	14.44	14.44
1.88–1.86	13.89	13.89	13.89	13.89	13.89	13.89	13.89	13.89	13.89	13.89
1.90–1.88	11.89	11.89	11.89	11.89	11.89	11.89	11.89	11.89	11.89	11.89
1.90–1.88	12.54	12.54	12.54	12.54	12.54	12.54	12.54	12.54	12.54	12.54
1.92–1.90	13.71	13.71	13.71	13.71	13.71	13.71	13.71	13.71	13.71	13.71
1.94–1.92	12.61	12.61	12.61	12.61	12.61	12.61	12.61	12.61	12.61	12.61
1.96–1.94	14.20	14.20	14.20	14.20	14.20	14.20	14.20	14.20	14.20	14.20
1.98–1.96	14.69	14.69	14.69	14.69	14.69	14.69	14.69	14.69	14.69	14.69
2.00–1.98	15.97	15.97	15.97	15.97	15.97	15.97	15.97	15.97	15.97	15.97
Mean	13.24	13.22	13.20	13.25	13.28	13.32	13.39	13.46	13.52	13.62
n	50	51	52	53	54	56	56	57	58	59
s	1.21	1.20	1.20	1.24	1.24	1.28	1.36	1.44	1.50	1.70
Mean + 2 s	15.66	15.63	15.60	15.73	15.76	15.89	16.11	16.33	16.51	17.03

Table 7.2 The statistical clustering of sample depths, core TCA9b, based on lead concentrations. The lowermost sample depth in the elevated lead cluster is taken to be the depth at which lead levels increase above background concentration. The single linkage algorithm excluded one sample, 0.76–0.74 m (2.67 μg g^{-1}), from membership of either cluster.

Clustering technique	'Background' lead group (μg g^{-1})			Elevated lead group (μg g^{-1})			Depth of initial elevated lead (m)
	n	*mean*	*s*	*n*	*mean*	*s*	
Hierarchical clustering (single linkage)	58	13.52	1.50	41	19.65	2.35	0.84–0.82
Hierarchical clustering (Ward's method)	58	13.52	1.50	42	19.73	2.39	0.84–0.82
K-means clustering	67	13.92	1.75	33	20.60	1.91	0.84–0.82

7.4 Palaeomagnetic dating of core TCA9b

7.4.1 Palaeomagnetic dating of recent sediments

Sediment sequences spanning the past 400 years may be dated directly by matching their palaeomagnetic secular variation signatures with historically documented geomagnetic field fluctuations (Thompson and Oldfield, 1986). Such sediment sequences must be either rapidly deposited, or sampled at a sufficiently high resolution to ensure that tell-tale magnetic features are recorded, and not averaged out as part of some longer term trend. The easiest historical field features to recognise are the turning points (maxima and minima) in inclination and declination. Thompson and Oldfield (1986) cited the 1810 AD westerly declination maximum and the 1710 AD inclination maximum observed in London, UK, as examples of good markers; Turner and Thompson (1979), for example, observed these distinct features in the uppermost sediments of Loch Lomond, UK. The dating of recent sediments by these methods has been rare, and in Australia no such studies have been attempted to date. This is probably due in part to the dearth of historical magnetic observations made prior to the 18th century across the continent against which a remanence record may be compared, as well as a lack of interest in sediments of historical age.

7.4.2 Palaeomagnetic dating studies from Australia

The most frequently cited Australian record, and that of most relevance to the eastern Australian region is the Southeast Australian mastercurve (KBG). This record of declination and inclination is composed of palaeomagnetic measurements from Lakes Keilambete, Bullenmerri and Gnotuk in southwest Victoria (Barton and Polach, 1980; Barton and McElhinney, 1981). The curve is 10 000 radiocarbon years long, but provides only limited information on secular changes over the last two millennia (20 data points, or an average resolution of only 100 years). The Keilambete record possess 10 magnetic values for the last 1000 years radiocarbon years, the Lake Bullenmerri curve has six, while that from Lake Gnotuk has none. The resolution of the KBG record is thus adequate for recording Holocene scale perturbations, but sub-millennial or sub-centennial information is lacking and the amplitude of such shifts is likely to be blurred into a much longer-term trend in the KBG record.

A second record for the eastern Australian region comes from core JA, from Lake Johnston, Tasmania (Anker *et al.*, 2001). This contains quite detailed information for much of the Holocene, but has no dated palaeomagnetic features less than 2000 BP old (dated by correlation with the KBG mastercurve), while the most recent direct date is 3340 ± 100 ^{14}C years BP. Anker *et al.* (2001) estimated that the top 700 mm of sediment is absent from core JA, thus explaining the absence of a late Holocene record of palaeomagnetic variations. Interestingly, their work led them to suggest that correlating palaeomagnetic records is best done in Australia using inclination records, which display characteristic swings (Anker *et al.*, 2001). This had first been noted by Barton and McElhinny (1981), whose palaeomagnetic records showed greater swings in inclination than in declination.

Finally, there are the geomagnetic secular variation records from Lake Barrine and Lake Eacham on the Atherton Tablelands in far north Queensland (Constable and McElhinny, 1985). The Lake Eacham record was derived by stacking directional data from four sediment cores, and dated by 11 ^{14}C dates which show that the sequence spans the last 5000 radiocarbon years. For the last 1000 years alone, the Eacham record has nearly 20 measures of inclination and declination. Although more thoroughly dated (n = 45), the Lake Barrine record has no information for the last 1600 years BP.

7.4.3 The eastern Australian Inclination Record (eAIR)

The existing information on secular geomagnetic variations is inadequate to allow the palaeomagnetic dating of sediments deposited in the last millennium. This lacuna has been filled by exploiting geomagnetic data from two separate sources. By combining all known inclination observations spanning the period from 1770 to 2003 (n = 89) in eastern Australia with a range of archaeomagnetic inclination measurements, a record of the change in magnetic inclination in the region has been developed (Table 7.3).

The eAIR record (Figure 7.7) has been standardised to the latitude and longitude of Sydney, New South Wales, eastern Australia. This has been done for three main reasons. First, the earliest direct observation of inclination taken in the Australia region was made in Sydney, by Captain James Cook in 1770. Secondly, measurements made beyond Sydney during the crucial period of the early 19th century (mainly in Tasmania) can be standardised to Sydney's location using the corrections devised by Barton and Barbetti (1982). Finally, it was necessary to standardise the observational record of inclination variation to a location not too distant from the Tocal Homestead Lagoon study site, so that direct comparisons could be made between it and the geomagnetic record preserved in the lake's sediments. In doing so, distinct palaeomagnetic features identifiable in both records could be correlated, and the corresponding features in the Tocal sediments allocated a date.

Records of magnetic inclination for the millennia before the 18th century were obtained from measurements of the palaeomagnetism of ancient fireplaces and burnt tree-stumps in eastern Australia by Barbetti (1977) (n = 12). Repeat measurements made on each fireplace allowed confidence limits to be assigned to inclination values for the lower part of the record. Carbon-14 dating of the fireplaces and tree stumps allowed them to be incorporated into the curve's chronology. Most of the magnetic measurements from eastern Australia in the 18th and 19th centuries were taken from *The Historical Magnetic Database* (Jonkers *et al.*, 2003). These data were compiled from the written observations of mariners (involved in trade and exploration) from the early 16th century onwards, and from previous compendia of magnetic observations by Hansteen (1819), Sabine (1872; 1875; 1877) and Bloxham (1986).

Wherever possible, these data were checked for accuracy against their original sources. The majority of the data from the beginning of the 20th century onwards was taken from the British Geological Survey's compilation *World-wide Magnetic Survey* (from their website). These data are from various stations in the Sydney region, all located within 150 km of Sydney GPO. Five very recent observations of inclination from the Newcastle region (-32.633°, 151.583°) were provided by Geoscience Australia (data from their website). The full list of data sources is shown in Table 7.3.

Table 7.3 Variations in magnetic inclination in eastern Australia from *c.* 5000 BC to 2003 AD based on archaeomagnetic measurements and direct observations. Inclinations in italics have been corrected to Sydney, New South Wales following the procedures of Barton and Barbetti (1982). Inclinations marked with an asterix are mean values. As no coordinates are provided by Barbetti (1977) for the locations of his archaeomagnetic measurements, the latitude and longitude of Willandra Lakes, New South Wales have been used instead. Radiocarbon ages were converted to calendar years using CalPal (the Cologne Radiocarbon Calibration and Paleoclimate Research Package).

Date	**Error (1 s)**	**Latitude**	**Longitude**	**Inclination**	**Error (1 s)**	**Source**
(years AD)	**(± years)**	**(°)**	**(°)**	**(°)**	**(°)**	
-5058	194	-33.717	143.000	-50.90*	0.68	Barbetti (1977)
-2893	285	-33.717	143.000	-55.50*		Barbetti (1977)
-2576	167	-33.717	143.000	-48.30*		Barbetti (1977)
-586	136	-33.717	143.000	-51.30*	1.71	Barbetti (1977)
-30	121	-33.717	143.000	-57.40*		Barbetti (1977)
613	96	-33.717	143.000	-61.46*	1.99	Barbetti (1977)
1082	112	-33.717	143.000	-50.79*	2.36	Barbetti (1977)
1217	135	-33.717	143.000	-48.20*	2.44	Barbetti (1977)
1253	49	-33.717	143.000	-49.76*	2.83	Barbetti (1977)
1271	55	-33.717	143.000	-54.30*		Barbetti (1977)
1527	77	-33.717	143.000	-61.70*	2.36	Barbetti (1977)
1688	115	-33.717	143.000	-67.40*	1.24	Barbetti (1977)
1770.38		-33.800	151.400	-67.00		Cook (1771)
1770.43		-34.000	151.400	-67.00		Cook (1771)
1777.08		-43.367	147.467	*-62.17*		Jonkers *et al.* (2003)
1777.08		-43.350	147.483	*-61.90*		Hansteen (1819)
1777.92		-43.683	147.333	*-62.90*		Jonkers *et al.* (2003)
1777.94		-43.350	147.550	*-62.87*		Jonkers *et al.* (2003)
1792.00		-43.600	147.000	*-63.50*		Jonkers *et al.* (2003)
1792.42		-43.533	146.933	*-62.73*		Hansteen (1819)
1792.42		-43.533	147.100	*-62.40*		Hansteen (1819)
1793.00		-43.567	146.933	*-62.15*		Jonkers *et al.* (2003)
1793.17		-43.567	146.933	*-64.27*		Hansteen (1819)
1824.50		-33.850	151.283	-62.30		Jonkers *et al.* (2003)
1831.50		-33.850	151.283	-62.85		Jonkers *et al.* (2003)
1831.50		-33.850	151.283	-62.85		Jonkers *et al.* (2003)
1836.50		-33.850	151.283	-62.82		Jonkers *et al.* (2003)
1836.50		-33.850	151.283	-62.82		Jonkers *et al.* (2003)
1838.50		-42.900	147.300	*-61.90*		Jonkers *et al.* (2003)
1840.50		-42.900	147.400	*-62.60*		Jonkers *et al.* (2003)
1841.68		-33.900	151.300	-62.80		Sabine (1844, p. 153)
1841.68		-33.933	151.000	-62.67		Sabine (1844, p. 153)
1845.50		-42.900	147.400	*-62.50*		Jonkers *et al.* (2003)
1860.00		-37.800	145.000	*-62.70*		Jonkers *et al.* (2003)
1880.00		-37.800	145.000	*-62.70*		Jonkers *et al.* (2003)
1890.00		-37.800	145.000	*-62.60*		Jonkers *et al.* (2003)
1896.22		-42.867	147.367	*-63.10*		Jonkers *et al.* (2003)
1896.29		-43.133	147.867	*-63.57*		Jonkers *et al.* (2003)
1896.45		-33.850	151.217	-62.83		Jonkers *et al.* (2003)
1897.22		-33.850	151.217	-62.90		Jonkers *et al.* (2003)
1897.29		-43.133	147.867	*-63.42*		Jonkers *et al.* (2003)
1897.38		-33.850	151.217	-62.83		Jonkers *et al.* (2003)
1897.38		-33.850	151.217	-62.83		Jonkers *et al.* (2003)
1897.38		-33.850	151.200	-62.78		Jonkers *et al.* (2003)
1897.85		-33.850	151.217	-62.90		Jonkers *et al.* (2003)
1906.90		-33.743	151.067	-63.08		British Geological Survey
1913.18		-33.742	151.067	-63.20		British Geological Survey
1913.24		-33.742	151.233	-63.28		British Geological Survey
1916.05		-32.748	151.067	-63.30		British Geological Survey
1916.09		-33.742	151.067	-63.29		British Geological Survey
1916.40		-33.742	151.067	-63.30		British Geological Survey
1921.80		-33.742	151.067	-63.48		British Geological Survey
1921.81		-33.742	151.067	-63.44		British Geological Survey
1922.85		-32.758	151.067	-63.48		British Geological Survey
1923.48		-33.742	151.067	-63.54		British Geological Survey
1936.34		-33.742	151.067	-63.60		British Geological Survey
1937.17		-33.763	150.917	-63.73		British Geological Survey
1937.66		-33.763	150.917	-63.68		British Geological Survey
1938.05		-33.763	150.917	-63.74		British Geological Survey
1939.33		-33.763	150.917	-63.70		British Geological Survey
1940.01		-33.763	150.917	-63.72		British Geological Survey
1940.26		-33.763	150.917	-63.83		British Geological Survey

1941.50		-33.763	150.917	-63.68		British Geological Survey
1942.07		-33.763	150.917	-63.69		British Geological Survey
1943.30		-33.742	151.283	-63.62		British Geological Survey
1943.39		-33.730	151.417	-63.40		British Geological Survey
1944.19		-32.760	151.267	-63.83		British Geological Survey
1944.21		-33.860	151.267	-63.77		British Geological Survey
1945.93		-33.610	150.725	-63.60		British Geological Survey
1947.02		-33.610	150.725	-63.63		British Geological Survey
1955.21		-33.617	150.733	-63.81		British Geological Survey
1957.50		-33.800	150.867	-64.07		British Geological Survey
1960.95		-32.750	152.000	-63.74		British Geological Survey
1960.95		-33.383	151.433	-64.10		British Geological Survey
1965.89		-33.767	151.883	-63.12		British Geological Survey
1969.08		-32.698	151.688	-63.21		British Geological Survey
1969.08		-32.828	151.598	-63.57		British Geological Survey
1969.08		-33.183	151.418	-63.77		British Geological Survey
1969.08		-33.316	151.251	-63.94		British Geological Survey
1969.08		-33.441	151.593	-63.44		British Geological Survey
1969.09		-33.880	151.963	-63.12		British Geological Survey
1969.09		-32.520	151.265	-63.08		British Geological Survey
1969.09		-32.653	151.453	-63.42		British Geological Survey
1969.34		-32.708	151.833	-63.20		British Geological Survey
1970.00		-32.797	151.130	-63.16		British Geological Survey
1970.00		-32.628	151.250	-63.12		British Geological Survey
1970.00		-32.698	151.453	-63.45		British Geological Survey
1970.00		-32.708	151.830	-63.27		British Geological Survey
1970.00		-32.800	151.688	-64.47		British Geological Survey
1970.00		-33.047	151.598	-64.32		British Geological Survey
1970.00		-33.183	151.418	-63.85		British Geological Survey
1970.00		-33.317	151.252	-64.32		British Geological Survey
1982.52		-33.482	151.836	-63.56		British Geological Survey
1985.00		-32.633	151.583	-63.39		Geoscience Australia
1985.31		-32.796	151.836	-63.55		British Geological Survey
1986.00		-32.633	151.583	-63.40		Geoscience Australia
1989.28		-32.796	151.836	-63.61		British Geological Survey
1992.25		-32.796	151.836	-63.59		British Geological Survey
2002.00		-32.633	151.583	-63.27		Geoscience Australia
2003.00		-32.633	151.583	-63.17		Geoscience Australia
2003.00		-32.633	151.583	-63.17		Geoscience Australia

7.4.4 The problem with La Pérouse

A preliminary compilation of eastern Australia magnetic observations showed a significant discrepancy for the value of inclination in Sydney in the 1770s. This period is of vital importance for the eastern Australian master curve, as it is the time of the very first direct measurements of inclination by European explorers. It is also close to the date of first European contact with the Australian environment, and thus the date of the stratigraphic level separating pre- and post-contact sedimentation. As discussed elsewhere, the difficulty in dating this period in Australian environmental history is perhaps the biggest challenge facing scientists working in this field today (section 3.6).

Captain James Cook made the first observation of magnetic inclination in eastern Australia while sailing north along the east coast on his first round the world voyage. While in Botany Bay on 25 April 1770, Cook recorded an inclination of -67° (Cook, 1771). He obtained the same reading in nearby Port Jackson on 6 May 1770 (Cook, 1771). Further measurements of inclination were not made in the central eastern Australian region until 1788, this time by the French explorer Jean-François de Galaup La Pérouse.

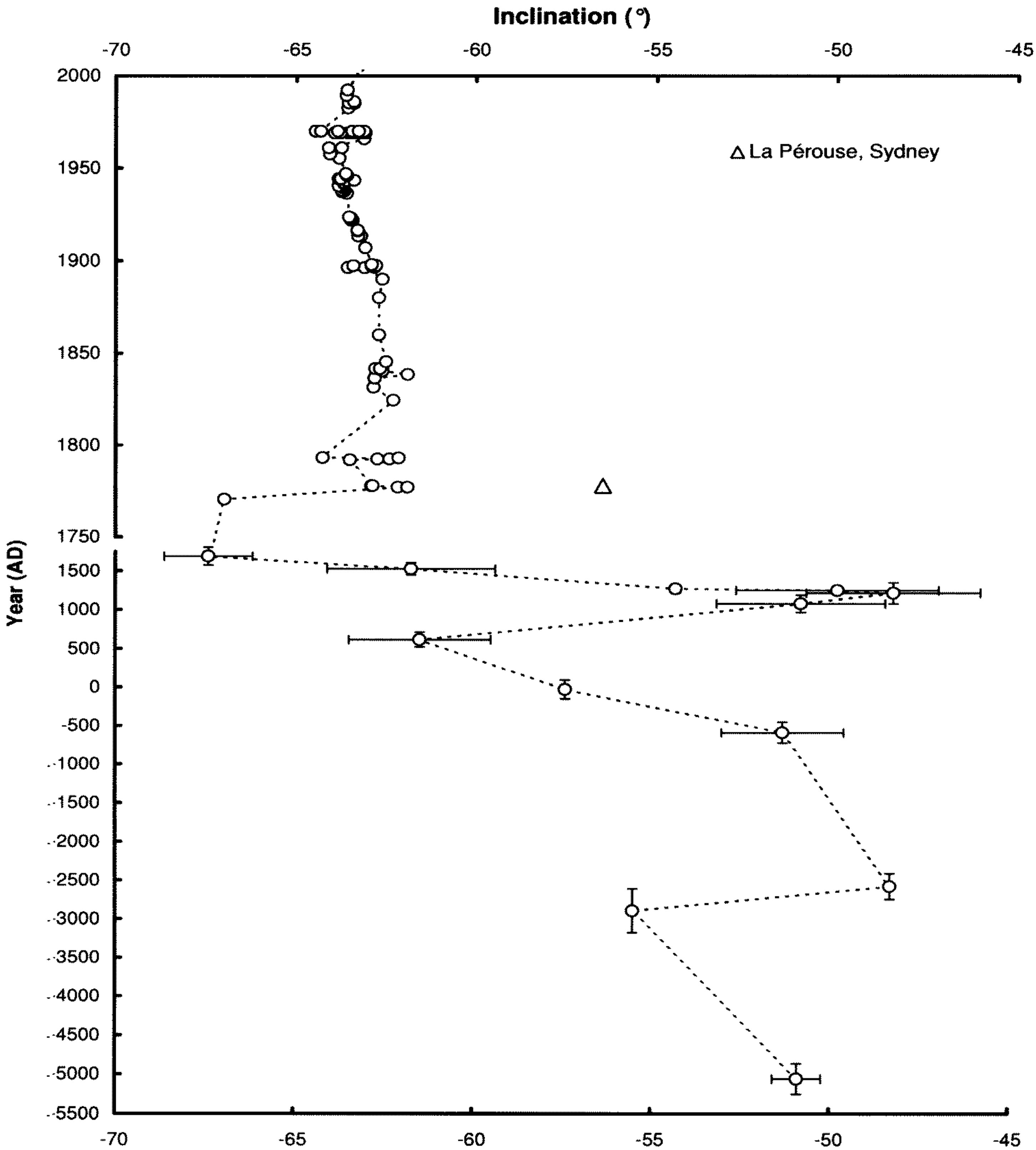


Figure 7.7 The eastern Australian inclination record (eAIR), based on historical observations of inclination since 1770 AD. Data earlier than this are based on the archaeomagnetic measurements made by Barbetti (1977). Horizontal and vertical error bars represent 66% confidence limits for each measurement.

On 22 January 1788, La Pérouse determined the inclination in Botany Bay to be -56°32' (Hansteen, 1819, p. 108), significantly shallower than the measurements of Cook six years earlier. Unfortunately the original written source of La Pérouse's magnetic observations is presently unknown, precluding any scrutiny of his original records, or any insight into the conditions under which they were executed. Assuming that the value of -56°32' recorded by Hansteen (1819) is correct, what explanation can be found to explain the discrepancy? Is either Cook or La Pérouse's measurement at fault? Had there been a rapid shift in field direction, or a magnetic storm? The disagreement between the inclination measurements of La Pérouse and those obtained by others from this period is, however, not an isolated occurrence.

Christopher Hansteen (1819, p. 148) noted that La Pérouse's observations of inclination were much shallower than those of other late 18th century observers. La Pérouse recorded an inclination of 28°0' in the Sandwich Islands (Hawaiian Islands, USA) (20°34' N, 156°5' W; off the northeast coast of the island of Maui) on 29 May 1786 (Hansteen, 1819). While offshore of the same islands on his third voyage, Vancouver recorded an inclination of 41°24' in March 1793 at Karakakooa, on the island of Hawaii (19°28' N, 155°56' W) (Vancouver, 1984). Cook determined the inclination south of the Hawaiian island of Kauai (21° 46', 159° 30') to be 42°37' in November 1788 (Cook *et al.*, 1782). Considering the very small changes in location and a time span of just five years, these measurements should not be greatly affected by the long-term drift of the magnetic field, and all three should be quite similar. Once again, La Pérouse's measurement is much shallower than the measurements of his contemporaries (Table 7.4). A final test of whether some fault lies in the recorded inclination observations of La Pérouse comes from a comparison of his measurements with the modelled inclination data for eastern Australia for the late 1700s. The modelled data from Jackson *et al.* (2000) (*gufm1* model) predicts the inclination in Sydney, Australia in 1778 to be -62.5299°, significantly different to La Pérouse's observation of -56°32'.

Given that La Pérouse's observations, as recorded by Hansteen (1819) (the only known source of these data), are consistently incorrect, this being noted at different locations and at different times, his single measurement made in eastern Australia was removed from the eAIR master curve.

Table 7.4 Measurements of inclination made in the Hawaiian Islands between 1786 and 1793.

Date	Voyage	Latitude	Longitude	Inclination	Source
29 March 1786	La Pérouse	20°34' N	156°5' W	28°0'	Hansteen (1819)
8 November 1788	Cook	21°46' N	159°30' W	42°37'	Cook *et al.* (1782)
March 1793	Vancouver	19°28' N	155°56' W	41°24'	Vancouver (1984)

It should be noted that Cook's measurement of -67° is somewhat steeper than the modelled inclination data of Jackson *et al.* (2000). Their predicted value, however, is based on a massive data set that includes La Pérouse's erroneous inclination value for the year 1778, skewing their model's output for this geographical location and time period. Cook's measurement, however, does plot between the most recent archaeomagnetic measurement made by Barbetti (1977) from 1688 (± 115) AD, and the next year for which data are available, 1777 (n = 4) (Figure 7.7). This segment of the master curve appears to record a relatively rapid increase in magnetic inclination which was completed by the end of the 18th century.

7.4.5 Other data excluded from the eAIR curve

Other than the curious measurement of La Pérouse, only a single other data point mined from the historical record was excluded from the eAIR curve. A measurement made at sea by the ship *The Pagoda*, some 500 km southwest of Tasmania in 1844 was not included (Jonkers *et al.*, 2003). This site was considered too remote to be included in the dataset, and adequate data for this period already exist for Sydney. For the later segment of the curve, values of magnetic inclination recorded at east Maitland, eastern New South Wales (-33.865°, 151.567°) were also removed from the dataset. A comparison of these observations with the nearby station of Newcastle, and with contemporaneous measurements from Sydney showed them to be consistently different by *c.* 1.5°. Once the east Maitland data were removed, sufficient data from other sites in the Sydney region provide satisfactory coverage for this period.

7.4.6 The 17th to 18th century inclination maximum in eastern Australia

In considering the last several centuries of the inclination record, the most striking feature is the inclination maximum recorded in the 17th and 18th centuries when inclinations reach their greatest value for the entire record: approximately -67.5° in 1688 ±115 AD. Although there are only a few data points for this segment of the curve, the data suggest that this period of very steep inclination may have lasted for over a century, and perhaps as long as two centuries. The first instrumental measurement of inclination in eastern Australia confirms that as late as 1770 the inclination in Sydney, New South Wales was still -67°. Inclination declined sharply by about 5° in the next 15 years, with a large number of measurements made in Tasmania (and corrected to Sydney) showing inclinations of the order of -62°. This was confirmed in 1824 when H.M.S. *Pandora* made magnetic measurements in Port Jackson, Sydney, recording an inclination of -62°18'.

7.4.7 Comparisons with other Australian records

The 17th–18th century eastern Australian inclination maximum appears to be replicated in several of the Australian secular magnetic records that cover this period. Core K1D from Lake Keilambete in southwest Victoria (Barton and Barbetti, 1982) shows inclination values steepening steadily after about 1500 years BP, reaching the maximum for the core at approximately 200 years BP. Sample resolution in this segment of the core does not allow any more precise information on the timing on this feature, however.

Although it is difficult to discern, the sediments from Lake Eacham (Constable and McElhinny, 1985) also appear to exhibit an inclination maximum (of nearly -80°) in the period 1600–1700 AD, with the two measurements above this recording a rapid decline in inclination towards that of the axial dipole (Constable and McElhinny, 1985, Figure 9b). Constable *et al.*'s (2000, Figure 3) Thiuer model of inclination changes worldwide from 1000 BC to 1800 AD is derived from a large number of archaeomagnetic inclination records, instrumental records and sediment records from sites around the world. Their model clearly shows the inclination anomaly identified in the eAIR. This began in the 1500s AD and reached its climax in 1600–1700 AD.

Although the excellent geomagnetic record from Lake Pounui in central New Zealand (Turner and Lillis, 1994) may be considered too far away to be directly comparable (the site lies approximately 2000 km east of Sydney), it is interesting to note that it too appears to preserve some evidence of this distinctive inclination anomaly. Inclination increases steadily from a local minimum after 500 years BP, reaching a maximum at about 1700 AD (though relatively large errors are attached to these inclination measurements at this point). It is difficult to isolate a clear turning point in the curve in the 18th century, however, as the upper most sediments in the Pounui core failed to yield stable NRMs (Turner and Lillis, 1994).

7.4.8 Core correlation, geomagnetic features and sediment dating

The inclination maximum identified from the historical and archaeometric record is clearly discernible in the core TCA9c record (Figure 7.8). This feature, centred on sediments at 2.77 m in core TCA9c, corresponds to 1.28–1.26 m in core TCA9b. Thus, the sediments deposited at 1.28–1.26 m depth in core TCA9b may be considered to date from between 1688±115 and 1770 AD. Although the error bars of the ^{14}C determined age suggests that the sample may date from as late as 1918 AD (at 95% confidence), inclination in Sydney had shifted quickly to *c.* -62° by as early as 1777 AD, and definitely by 1793 AD (Table 7.3 and section 7.4.6). Thus the palaeomagnetic age may be constrained, with its upper age limit being reset to 1777 AD.

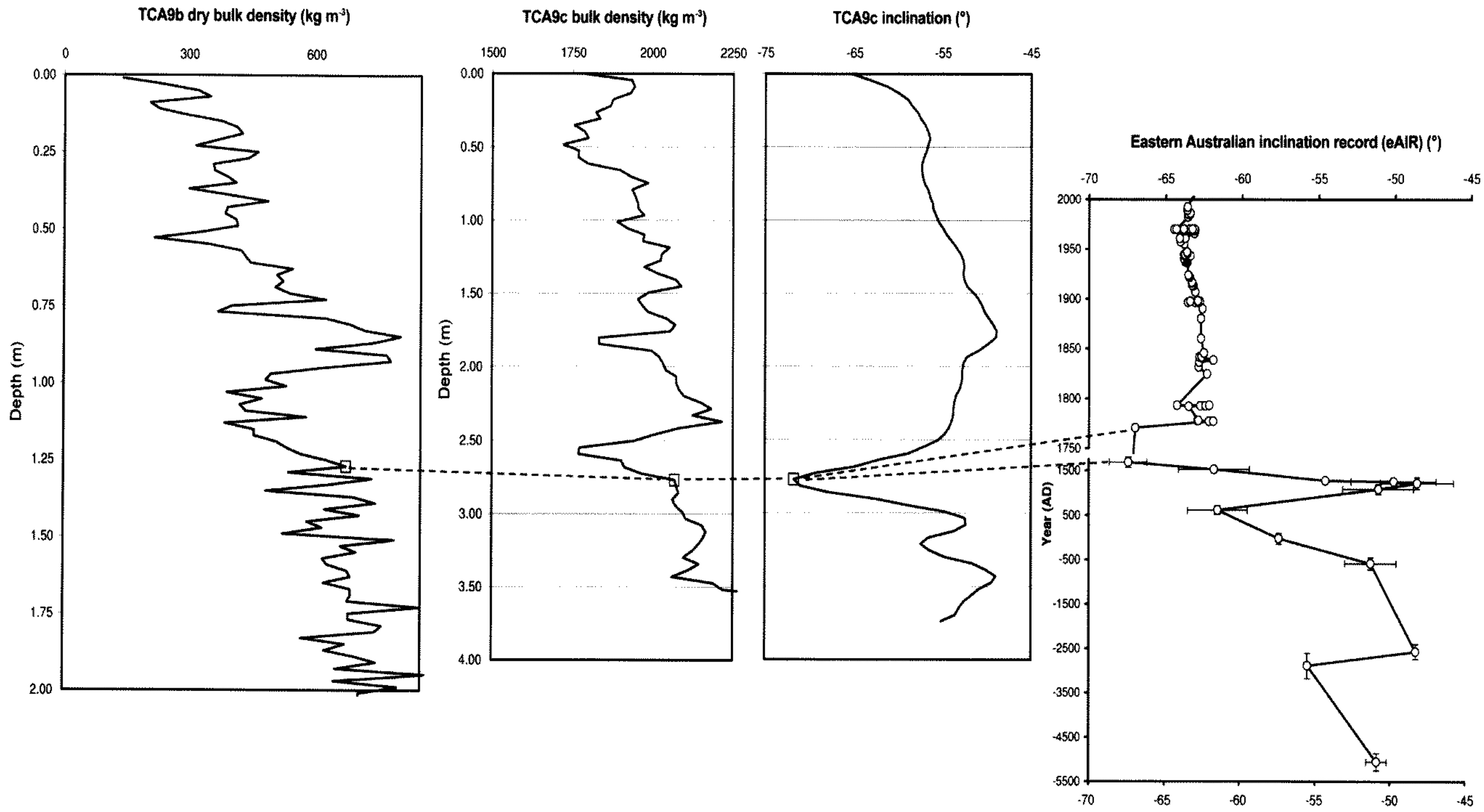


Figure 7.8 The correlation of the palaeomagnetic record from core TCA9c with core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia, via the correlation of downcore patterns in sediment bulk density. Difficulties in calibration resulted in the bulk density measurements of core TCA9c being approximately double those determined for core TCA9b. However, the downcore pattern remains unaffected.

7.5 Carbon-14 chronology

7.5.1 Carbon-14 ages from core TCA9b

The radiocarbon data obtained from five samples from core TCA9b are presented in section 6.6.1, and are reproduced here for convenience in Table 7.5 and Figure 7.9. These results show that the 2.00 m sequence from core TCA9b contains a sedimentary record spanning the last 2000 years.

7.5.2 Palaeomagnetic dating versus ^{14}C ages

The palaeomagnetic date ascribed to the sediments from 1.28–1.26 m in core TCA9b is statistically identical to two AMS ^{14}C dates from this depth (Table 7.5), showing that these sediments cannot be older than the 14th century AD. Curiously, the next sample up the sequence (1.26–1.24 m) was dated by AMS ^{14}C methods to the 11th century AD. The evidence from the palaeomagnetic date alone suggests that this is unlikely, unless significant disturbance of the sediment sequence had occurred. As discussed in section 6.6.1, there is evidence that this is unlikely. The ^{14}C date from 1.26–1.24 m was made on microscopic material which is more easily reworked than macroscopic charcoal, and thus more likely to yield anomalously old dates. It is therefore thought to represent the age of much older microscopic organic matter that has been reworked from the catchment and transported into the lake (probably during a phase of increased catchment erosion), rather than the age of the stratigraphic level from which it was obtained. Evidence from other Australian sites suggests that the impacts of European land use are such that reworking of charcoal is likely to be a problem from the time of European contact onwards.

There are three main lines of evidence that suggest that no significant delays in the transport of allochthonous macroscopic charcoal to the lake have occurred elsewhere in the core TCA9b sediment sequence. Importantly, this means that the ^{14}C chronology from core TCA9b below 1.26 m depth may be interpreted with confidence. First, with the exception of the date from 1.26–1.24 m, all the ^{14}C ages from core TCA9b increase with sediment depth. No other age inversions are recorded; age inversions in the sequence would be expected if significant and inconsistent delays in the transport of charcoal to the lake had occurred. Secondly, this studies use of duplicate ^{14}C determinations from a single depth (1.28–1.26 m) has shown that two separate charcoal samples from the same depth have identical ages. Although this may be the result of two separate macro-charcoal fragments being delayed in an identical fashion, this is an unlikely scenario. Finally, the independent dating of the sequence by palaeomagnetic methods has shown the age of the sediments from 1.28–1.26 m, as determined by AMS ^{14}C dating of allochthonous macro-charcoal, to be correct (section 7.5.2).

Table 7.5 Radiocarbon ages and calibrated calendar year determinations made on charcoal and organic extracts from core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Calibrations were performed using CalPal (the Cologne Radiocarbon Calibration and Paleoclimate Research Package).

Sample Depth (m)	Laboratory Code	^{14}C age (^{14}C years)	Error (^{14}C years)	Mean calendar age (calibrated years BP)	Error (1 s) (calibrated years BP)	Median calendar date (years AD)	Median calendar date error (1 s) (years AD)
1.26–1.24	Beta-190236	960	40	864	53	1086	53
1.28–1.26	Beta-183139	570	40	586	40	1364	40
1.28–1.26	Wk-8573	430	70	435	81	1515	81
1.32–1.30	Wk-7707	730	70	661	64	1289	64
1.94–1.92	OZE079	2090	60	2065	74	-115	74

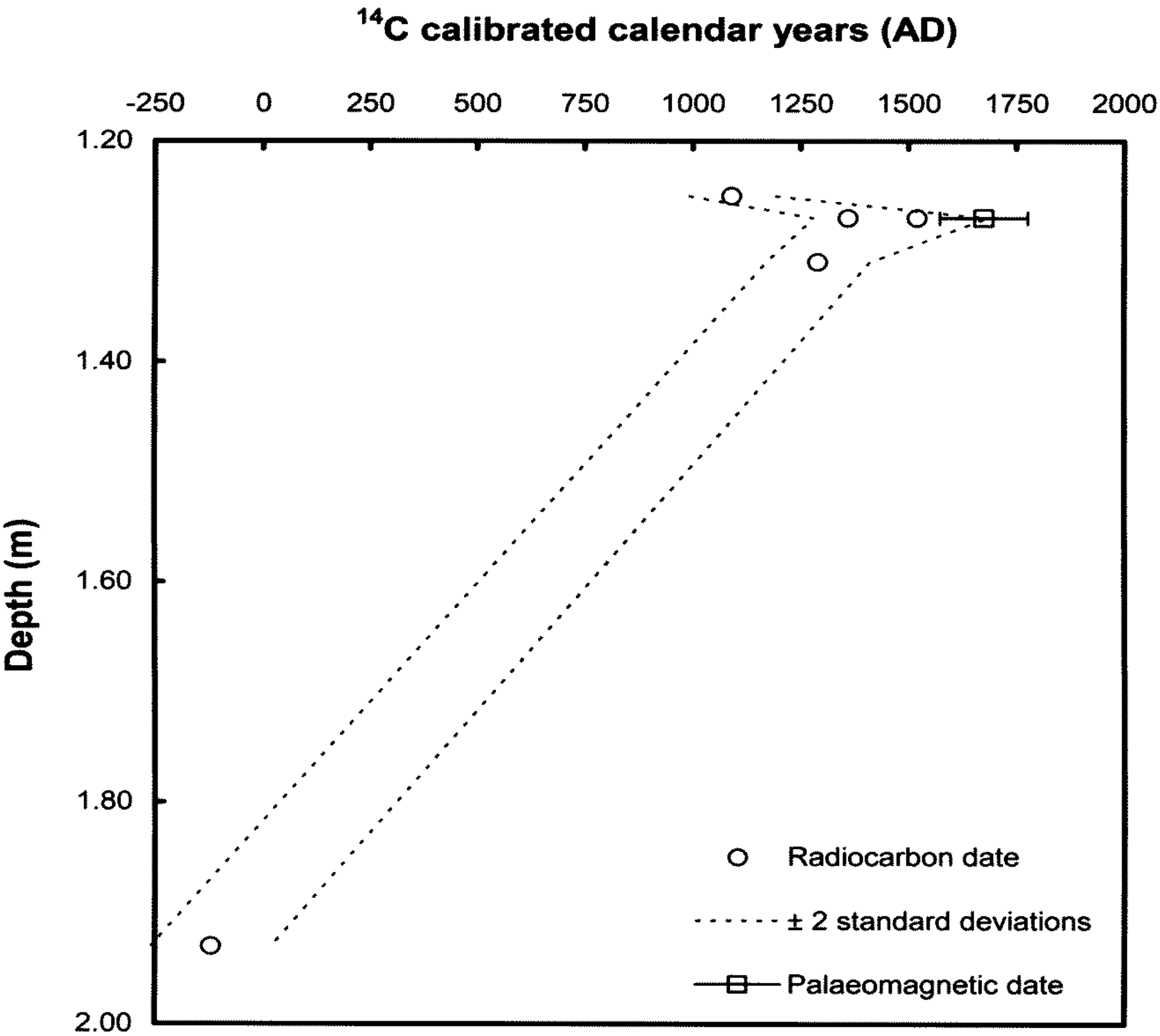


Figure 7.9 The AMS ^{14}C chronology and palaeomagnetic date from core TCA9b, Tocal Homestead Lagoon, eastern Australia. Broken lines denote envelopes of 95% confidence for the ^{14}C age determinations. Horizontal error bars on the palaeomagnetic date represent 68% confidence limits.

7.6 Determination of ^{210}Pb chronologies

(from Cook *et al.*, in press)

7.6.1 Lead-210 dating in Australia

The modelling of unsupported ^{210}Pb profiles to date sedimentary sequences was first described by Goldberg (1963) and first applied to dating lake sediments by Krishnaswamy *et al.* (1971). Only a summary of its use in dating sediment records in Australia is presented here. With recent studies extending the range of ^{210}Pb dating to *c.* 180 years in Australia through the use of careful model calibration and parallel dating programs (Harle *et al.*, 2002; Gale *et al.*, 2004), ^{210}Pb has arguably become the most successful and widespread means of dating environmental records of the post-contact era on the continent.

A growing corpus of research has now successfully used ^{210}Pb dating to provide information on environmental change during the historical period in Australia. It should be noted, however, that only a handful of these studies has calibrated the models used in the determination of sediment ages (Gale *et al.*, 1995 and later papers), a vital step in deriving ^{210}Pb chronologies. Without testing the modelling of ^{210}Pb ages, any predicted dates cannot be accepted with any confidence (Smith, 2001; Abril, 2004). Perhaps of even more concern have been those studies that have simply used the detection of excess ^{210}Pb at the lowest depth in a sequence as a rough time-marker of, what seems to be an arbitrary figure, *c.* 120 years (Boon and Dodson, 1992; Dodson *et al.*, 1993).

Smith (1982) used ^{210}Pb methods to date cores from Lake Tali Karng in the Victorian Alps (see also Smith and Hamilton [1985]), and in Melbourne and on the Baw Baw Plateau in southeast Victoria (1992). Gell *et al.* (1993) dated a sediment record of vegetation changes from Tea Tree Swamp in southeast Victoria. Gale *et al.* (1995) used ^{210}Pb methods to date sediments to two decades before the onset of official European settlement at Little Llangothlin Lagoon on the New England Tablelands of New South Wales. Haworth *et al.* (1999) established a ^{210}Pb chronology for Black Mountain Lagoon, on the New England Tablelands of New South Wales, that extended back 80 years. A short (*c.* 0.20 m) sediment sequence from Lake Barine, on the Atherton Tablelands of north Queensland was dated by Walker *et al.* (2000) using ^{210}Pb methods. Herczeg *et al.* (2001) employed ^{210}Pb methods to date sediments from Lake Alexandrina, South Australia, providing a 120 year record of nitrogen and carbon cycling. Matthai *et al.* (2001) used ^{210}Pb and ^{137}Cs dating to estimate sedimentation rates on the middle continental shelf, adjacent to Sydney, New South Wales. Brunskill *et al.* (2002b) estimated rates of sedimentation and terrestrial and marine carbon burial on the Great Barrier Reef continental shelf, north Queensland. Fluvial sediment sequences near Sydney have been dated using ^{210}Pb methods by Harrison *et al.* (2003), providing the timeline for a pollution history near Sydney's main water supply. Seen *et al.* (2004) used ^{210}Pb methods to date sediments from the Tamar Estuary in Tasmania, providing the chronology for a Pb isotope pollution history. However, clear evidence of contamination of the sequence may have hampered the reliability of their ^{210}Pb chronology. A short ^{210}Pb chronology has been recently developed for a diatom based record of past water quality from the floodplain of the Snowy River (MacGregor *et al.*, 2005). However, the coarse sampling interval used in this study hampered the resolution of the chronology. The top 0.22 m of a Holocene fossil pollen and charcoal record from the Fleurieu Peninsula, South Australia has been dated by ^{210}Pb by Bickford and Gell (2005). Other recent studies include that of Leahy *et al.* (2005), who used ^{210}Pb methods in an attempt to derive a record of European impact from a floodplain lake near the city of Melbourne, Victoria.

It should also be noted that some attempts to apply the dating technique to lake systems in Australian have been unsuccessful. The downcore profile of excess ^{210}Pb in Burrinjuck Reserve, near Canberra, southeastern Australia showed considerable variation, leading to difficulties in modelling and age determination (Wasson *et al.*, 1987).

7.6.2 Farewell to the record of first European impact

A problem related to dating the impact of European settlement in Australia that should be mentioned here is that of distinguishing those sediments that were deposited during the period of first settlement. In most regions of Australia, European settlement took place at the end of the 18th century and in the first half of the 19th century, or 150–220 years ago. Given the timing of European contact, the main chronological technique employed by researchers in Australia has been ^{210}Pb dating. The ^{210}Pb dating technique is considered to be most useful in dating 'recent' sediments, those deposited within the last 130–150 years (Abril, 2004). It is the method of most relevance to studies of the impact of European settlement on the Australian environment. Several studies in Australia have shown it possible to use the technique to extend sediment chronologies to over 180 years (Gale *et al.*, 1995; Harle *et al.*, 2002), depending particularly on the initial concentrations of excess ^{210}Pb in the sediments (Gale *et al.*, 2004). In the approaching decades, as the excess ^{210}Pb of sediments contemporaneous with European settlement decays to levels beyond detection, the ^{210}Pb dating technique will no longer be capable of dating those sediments deposited at this time. This will be particularly problematic in those areas of the continent that were first settled over 150 years ago (most of southeastern Australia). At present there exists no other suitable technique that can date sediments of this age.

The only likely contender is Si-isotope dating, although this is capable of yielding dates covering the last 1000 years of environmental change (Nijampurka *et al.*, 1988), sediments must be extremely rich in diatoms (or some other contemporaneous source of silica), a condition rarely met in the Australian terrestrial environment. The technique is also significantly more costly and time-consuming than the ^{210}Pb procedure. The scientific community in Australia is soon to lose one of the best means of dating environmental change from the 19th century and future research in this field should take the rapidly dwindling dating potential of ^{210}Pb into consideration.

7.6.3 Modelling ^{210}Pb activities

Profiles of unsupported ^{210}Pb activity in sediment sequences may be used to establish chronologies by the use of one or more models. The two most widely employed models are the Constant Initial Concentration (CIC) and the Constant Rate of Supply (CRS) models. Others, such as the Constant Rate of Accumulation (CRA) model are inappropriate in systems in which the rate of sediment accumulation has varied significantly over time (Robbins and Herche, 1993, p. 219). Other models have been developed more recently, and are discussed by Abril (2004). Both the CIC and CRS models assume a constant input of unsupported ^{210}Pb to the lake and a constant residence time of unsupported ^{210}Pb in the water column; in other words, a constant flux of excess ^{210}Pb to the sediment. Both models also presume that there is no migration of ^{210}Pb within the sediments and no loss from the sediment to the lake water. In addition, the CIC model assumes that, at each stage of sediment accumulation, the initial concentration of unsupported ^{210}Pb is constant, regardless of any change in the rate of bulk sediment accumulation. The implication of this is that either the sedimentation rate must remain constant or the flux of unsupported ^{210}Pb must change in sympathy with changes in sediment accumulation. The second of these points clearly conflicts with the initial assumption of a constant flux of unsupported ^{210}Pb to the sediment.

Despite its inconsistent logic, the CIC model has been widely employed and it is probably the most frequently used means of deriving ^{210}Pb chronologies. By contrast, the CRS model, originally proposed by Goldberg (1963), and developed by Appleby and Oldfield (1978), assumes a constant flux of unsupported ^{210}Pb to the sediment, irrespective of changes in bulk sedimentation rates. Where it has been possible to check the results of the two models using other methods of dating, the CRS model appears to give the better results (Oldfield *et al.*, 1978; 1980; Appleby *et al.*, 1979; Battarbee *et al.*, 1980; Gale *et al.*, 1995). However, when the same set of ^{210}Pb data can be explained by different models, and thus lead to potentially different chronologies, problems arise. Smith (2001), addressing these problems, proposed a set of guidelines for those journals that publish sediment chronologies based on the ^{210}Pb method, 'The ^{210}Pb geochronology must be validated using at least one independent tracer which separately provides an unambiguous time-stratigraphic horizon.' Therefore, the use of independent time markers is mandatory in developing reliable ^{210}Pb-based chronologies of sedimentation.

7.6.4 Age calculations

The CIC model assumes that each layer in a sedimentary sequence had the same unsupported ^{210}Pb activity at the time of its deposition (although unsupported ^{210}Pb may vary between layers as a function of texture, or organic content, for example). Since ^{210}Pb decays logarithmically with time with a decay constant of 0.03114 a^{-1} (Höhndorf, 1969), the sediment age at any depth is given by:

Eqn. 7.1 $t_z = (1/k)\ \log_e(A_0/A)$

Where t_z = age at depth z, k = ^{210}Pb decay constant, A_0 = initial unsupported ^{210}Pb activity per unit mass and A = unsupported ^{210}Pb activity per unit mass at depth z.

The CRS model assumes a constant flux of unsupported ^{210}Pb to the sediment system, but allows the rate of sedimentation to vary over time. The sediment age at any depth is therefore given by the ratio of the unsupported ^{210}Pb activity below that point to the total unsupported ^{210}Pb activity in the sediment column:

Eqn. 7.2 $t_z = (1/k)\ \log_e(A_{total}/A_z)$

Where t_z = age at depth z, k = ^{210}Pb decay constant, A_{total} = total unsupported ^{210}Pb activity in the sediment column and A_z = unsupported ^{210}Pb activity below depth z.

1. CIC model: ^{210}Pb activities may vary as a function of sediment composition. Activities may be diluted by the addition of plant organic matter or coarse clastic material to the sediment, or altered by variations in bulk density. Accordingly, activities from core TCA9b were normalised against mineral mass, dry bulk density and mineral bulk density. It was not possible to normalise activities against variations in particle-size distribution, but there is little variation in sediment texture downcore. The deposits are almost uniformly composed of silt, occasionally containing rare or very rare grains of fine sand. The upper 0.10 m consists of silty clay. The strongest relationship with depth was obtained with activity per unit total mass ($\log_e A = 3.981 - 3.134\ z$, $r^2 = 0.71$, $n = 42$). This relationship was therefore used to estimate A_0 in the CIC model. The value obtained was substituted in equation (1) to determine t_z for each sample analysed. The error in the estimation of A_0

was determined from the standard error of the estimate of A in the regression equation, taking the conservative approach of employing the higher (positive) error after transformation back to arithmetic values.

2. CRS model: the CRS model calculates ages on the basis of the total activity of unsupported ^{210}Pb in the sediment column. Measurements of the dry bulk density of the sediments were therefore used to calculate the dry mass and thence the total activity of each sediment slice. Even with a very close sampling interval, such as that employed here, it is almost invariably the case that some gaps will occur where analyses have not been made. Analysts have therefore adopted several different procedures to estimate the activity of missing samples (e.g., Appleby [2001, p. 192]). Two approaches have been chosen here. First, the depth–activity regression relationship ($\log_e A = 3.981 - 3.134\ z$, $r^2 = 0.71$, $n = 42$) was employed to predict the missing values and their errors. Secondly, missing values were calculated as the simple arithmetic mean of the two adjacent known values. In both cases, uncertainties were estimated by general error propagation. The chronologies obtained using these two methods lie well within the ±1 standard deviation errors associated with each date, suggesting that the choice of interpolation procedure has little influence on the resultant chronology.

In addition, the activities need to be extrapolated towards the base of the core, to the point at which total ^{210}Pb activity is in radioactive equilibrium with the supporting ^{226}Ra. Not only is this difficult to do in practice, because of the errors associated with low activities, but erroneous values of A_{total} can result in highly inaccurate dates, particularly in the lower part of the core where cumulative activities are low. Therefore four different procedures were employed to estimate that part of the unsupported ^{210}Pb inventory below the lowest laboratory determination (at 1.26–1.24 m depth).

First, the depth versus $\log_e {}^{210}Pb_{excess}$ relationship was extrapolated for the entire core. In the first instance (model 1), values were determined to an arbitrarily selected depth of 1.40 m. Secondly, we used the same approach, but employed an iterative procedure to adjust the extrapolated cumulative activity at the base of the core so that it accorded with an independently dated reference level within this part of the sequence (model 2). The reference level that we chose occurs at 1.28 m, which an array of independent indicators mark as the first evidence for European activity at the site (section 6.4). Thirdly, we extrapolated values below the lowest measured activity by calculating the simple arithmetic mean of the adjacent known values (model 3).

Fourthly, the activities of the lowest four samples, which appeared to be declining exponentially with depth (Fig. 6.24), were used to estimate activities in the lower part of the sequence ($\log_e A$ = 91.19 - 75.07 z, r^2 = 0.82, n = 4). Values were determined to an arbitrarily selected depth of 1.40 m, although the depth of extrapolation had only a trivial effect on the chronology derived on the basis of measured activities (model 4).

It is essential that the uncertainty accompanying each age determination is established, though this is overlooked in many geochronological applications of ^{210}Pb analysis. The dates obtained here are expressed with an uncertainty of ±1 standard deviation. This includes the error involved in measuring isotopic activity, in determining the half-life of ^{210}Pb, in modelling variations in unsupported ^{210}Pb activity with depth in the sediment column and the propagation of each of these errors at each stage of the calculation. The results of the age determinations made using the five methods described above are plotted in Figure 7.10.

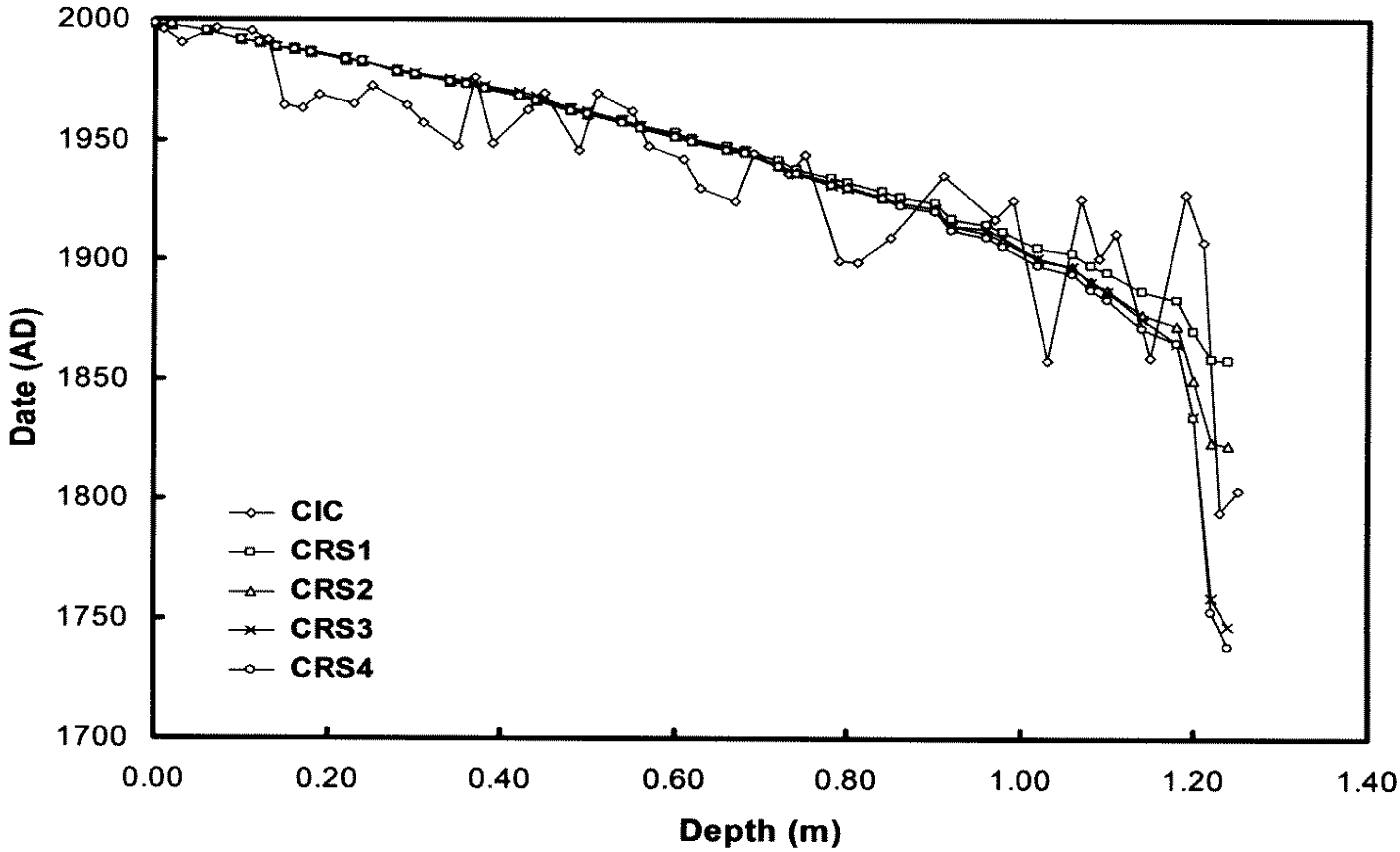


Figure 7.10 Lead-210 chronologies for core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia modelled using the procedures described in section 7.6.4. Note that, for clarity, the uncertainties associated with each date have been omitted from this plot. From Cook *et al.* (in press).

7.6.5 The flux of unsupported ^{210}Pb to the lake basin

Before the results of ^{210}Pb analyses are used to determine ages or sedimentation rates, it is important to establish whether there have been significant variations in the flux of unsupported ^{210}Pb to the system. These may arise in particular as a result of the inwash of material from the catchment to the lake. A crude means of assessing such variation is to compare the mean flux of unsupported ^{210}Pb to the sediments with the measured direct atmospheric fallout at the site (Oldfield and Appleby, 1984). It has not been conclusively demonstrated that there is a direct relationship between input to the sediments and atmospheric flux (and Robbins [1978, p. 371] has shown that the flux at any one point in the lake does not necessarily correspond with that expected from atmospheric input). Nevertheless, Oldfield and Appleby (1984, p. 120) have pointed out that where the flux to the sediment is much less than the measured atmospheric flux, results have often been poor, perhaps indicating sediment focusing or a low residence time of the lake waters.

The mean flux of unsupported ^{210}Pb to the deposits may be determined from:

Eqn. 7.3 $$F = A_0 S$$

where F = flux of unsupported ^{210}Pb per unit area, A_0 = initial unsupported ^{210}Pb activity per unit mass of sediment and S = mass accumulation rate of sediment per unit area.

The calculated mean flux for the time since European contact at site TCA9 in Tocal Homestead Lagoon is 171 Bq $m^{-2}a^{-1}$. Comparative records are hard to come by. Table 7.6 presents data from the three sites closest to the lagoon that were monitored by the Commonwealth X-ray and Radium Laboratory in the period 1964–1970. The calculated flux is of the order of twice the maximum recorded rate of deposition at these sites. There are several possible explanations for this difference. The high flux may be a consequence of unsupported ^{210}Pb carried into the lake in solution in catchment runoff. A second possibility is that much of the unsupported ^{210}Pb is derived from the erosion of catchment soils. Thirdly, ^{222}Rn is delivered to lake waters by diffusion from underlying sediments and from the decay of ^{226}Ra in the water column and in the waters of inflowing streams. Some radon is lost by diffusion across the lake surface, but the remainder decays in the water column to ^{210}Pb (Oldfield and Appleby, 1984, p. 95). Finally, Robbins and Edgington (1975, pp. 301–302) and Robbins (1978, p. 371) have suggested that high fluxes at particular sites in lakes are caused by the selective deposition of particles that scavenge ^{210}Pb from the water column.

Table 7.6 Deposition of ^{210}Pb in coastal central eastern Australia, 1964–1970, from Bonnyman and Molina-Ramos (*c.* 1971).

Site	Period of measurement	Annual deposition of ^{210}Pb (Bq $m^{-2}a^{-1}$)		
		Mean	**Minimum**	**Maximum**
Brisbane	1964–1970	63	50	73
Sydney	1964–1970	53	29	75
Berry	1965–1970	67	47	88

The first three of these possibilities are usually regarded as capable of contributing only small amounts of unsupported ^{210}Pb to the lake sediment system (though in all cases exceptions to this generalisation are known) (Gale *et al.*, 1995, pp. 395–396). Given the small extent and the ephemeral nature of the channels flowing into the lake, it is hard to see how the contribution of dissolved ^{210}Pb in runoff would be capable of increasing the flux to the lake sediments to approximately twice that of the atmospheric input. Similarly, the contribution of unsupported ^{210}Pb from the decay of ^{222}Rn in lake and contributing stream waters is generally very low, although Oldfield and Appleby (1984, p. 95) reported a single example in a relatively unusual environment in which riverine input and *in situ* production of ^{210}Pb are of the same order of magnitude as direct atmospheric flux. On the other hand, it is highly likely that material has been eroded into the lake basin as a result of the operation of European agricultural practices over the last 180 years, carrying with it atmospheric ^{210}Pb bound to soil particles. If this were the case, the flux to the lake should change over time in accordance with either change in the rate of erosion or with the first stripping of excess ^{210}Pb-rich topsoils from the catchment.

Reconstructing the history of ^{210}Pb flux is not straightforward. A reliable chronology is required; first to correct the measured values of $^{210}Pb_{excess}$ activity for decay since the time of deposition and secondly to calculate changes in sediment accumulation rates over time. If the chronology is in error, then both the calculated initial activities and the calculated accumulation rates will also be wrong. The CRS2 model has been employed here. Some assurance this is not grossly incorrect comes from section 7.7, Even so, the problems inherent in using an unproven timescale as the basis of these calculations are acknowledged. On the other hand, it would require great changes to the chronology to alter the calculated values sufficiently to draw conclusions that differ from those outlined below. The reconstructed history of unsupported ^{210}Pb flux to site TCA9 is shown in Figure 7.11.

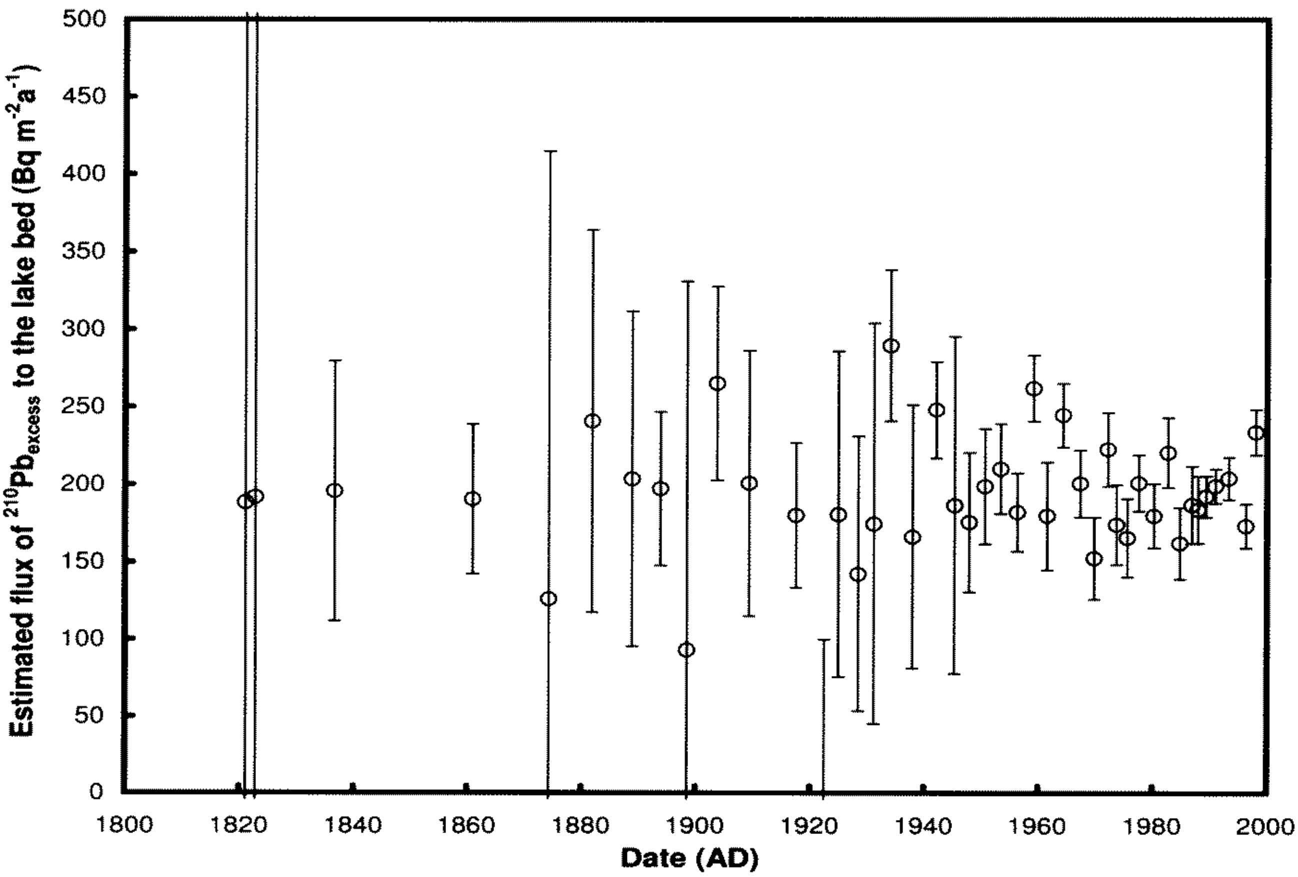


Figure 7.11 Estimated flux of $^{210}Pb_{excess}$ at site TCA9, Tocal Homestead Lagoon, Paterson, eastern Australia, *c.* 1821–1998 AD. The CRS2 chronology (see section 7.6.4) is used in the calculations. Sediment accumulation rates are obtained from continuous measurements of dry bulk density made down the core. The two samples displaying negative $^{210}Pb_{excess}$ activities are not plotted. The fluxes are expressed with an uncertainty of ±1 standard deviation calculated by general error propagation. From Cook *et al.* (in press).

The calculated fluxes are relatively constant over time. Nevertheless, the values are much greater than measured atmospheric fluxes in central eastern Australia; even the lowest of the mean determinations is greater than the highest instrumental measure. This would appear to rule out atmospheric flux as the sole contributor to unsupported ^{210}Pb in the sediments. On the other hand, there is no relation between variations in flux over time and the rates of lake sedimentation determined later in this thesis (section 8.4). Nor is there evidence of high flux at contact, indicative of the stripping of $^{210}Pb_{excess}$-rich topsoils into the basin. Significant amounts of unsupported ^{210}Pb are therefore unlikely to have been derived from catchment erosion. The most likely cause of the high flux at site TCA9 is thus the selective deposition of particles that scavenge ^{210}Pb from the water column. Nevertheless, irrespective of whether this or another mechanism operates, the relatively constant flux over time still meets the fundamental assumptions of the method. Indeed, Appleby and Oldfield (1992, p. 755) have noted that, even where non-atmospheric sources are significant, so long as there is a steady supply of unsupported ^{210}Pb to the sediment, the dating technique may still be valid. In addition they have noted that the CRS model is relatively unaffected by perturbations in ^{210}Pb flux (Appleby and Oldfield, 1992, p. 755).

7.7 Calibration of ^{210}Pb age models by independent methods

7.7.1 Introduction

Five models of the downcore pattern of excess ^{210}Pb activity in core TCA9b have been presented (Figure 7.10). The modelled chronologies are statistically indistinguishable (at ±1 standard deviation) down to 1.15 m. Using information from the ^{137}Cs, lead pollution and palaeomagnetic chronologies independently derived from core TCA9b, the most accurate ^{210}Pb age model may be identified.

7.7.2 Caesium-137 time horizon versus ^{210}Pb chronologies

The first appearance of ^{137}Cs in core TCA9b at 0.44–0.42 m (section 7.2) suggests that these sediments could not have been deposited any earlier than the late 1950s. At one standard deviation, this date is equivalent to the ages estimated by the CRS1, CRS2, CRS3 and CRS4 models (Figure 8.9). The ^{137}Cs time marker is also in agreement with the age estimate of the CIC model at 2 standard deviations.

7.7.3 Lead pollution history versus ^{210}Pb chronologies

The sediments from 0.84–0.82 m in core TCA9b preserve the first elevated lead levels at Tocal. Based on the history of lead pollution in Australia and at Tocal, these sediments were most probably deposited no earlier than 1932 (Cook and Gale, 2005). This independent date is identical to the ^{210}Pb predicted ages from the CRS1, CRS2, CRS3 and CRS4 models. At this depth (and at many others), the CIC model diverges distinctively from the four CRS models, and produces an estimated date of approximately 1900 AD, some three decades earlier than the independent time-horizon provided by the lead pollution stratigraphy. The CIC model is therefore rejected as representing the age–depth profile of core TCA9b.

7.7.4 Palaeomagnetic dating versus ^{210}Pb chronologies

Most of the divergence in the four remaining models takes place below 1.15 m. The inclination maximum identified in the historic and archaeomagnetic record from eastern Australia (section 7.4.6) and in the geomagnetic variation record from core TCA9b (section 7.4.8) provides the most solid information on the age of sediments in this segment of the core. Based on the palaeomagnetic dating undertaken in this study, the sediments from 1.28 –1.26 m in core TCA9b were deposited no later than 1770 AD. The four remaining ^{210}Pb age models extend to the stratigraphically lowest sample that contained excess ^{210}Pb, at 1.26–1.24 m. As these sediments directly overlie sediments from 1770 (maximum age), they cannot be any older than this.

For the sediments from 1.26–1.24 m, the four models predict the following mean ages: CRS1 model predicts an age of 1858 AD, the CRS2 model 1822 AD, the CRS3 model 1747 AD, and the CRS4 model 1738 AD. The CRS2 model displays the best fit with the palaeomagnetic evidence. Given that this model yields dates that closely conform with those provided by the ^{137}Cs and lead pollution markers, the CRS2 chronology was selected as the best representation of the age–depth curve for the top 1.26 m of the master core. The chronology produced using the CRS2 model is given in Table 7.7.

Table 7.7 Lead-210 dates as a function of depth in core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. The dates are derived from the CRS2 model.

Depth (m)	Date (AD)	Error (1 s)
0.00	1998.46	0.00
0.02	1997.73	6.69
0.06	1994.96	6.39
0.10	1991.74	6.47
0.12	1990.28	6.61
0.14	1988.49	6.80
0.16	1987.47	6.91
0.18	1986.35	7.04
0.22	1983.22	6.87
0.24	1982.21	6.98
0.28	1978.28	6.73
0.30	1976.94	6.87
0.34	1974.19	6.65
0.36	1973.16	6.76
0.38	1971.29	6.96
0.42	1968.28	6.61
0.44	1966.36	6.80
0.48	1962.26	6.77
0.50	1960.84	6.92
0.54	1957.48	7.19
0.56	1955.14	7.47
0.60	1951.58	7.52
0.62	1949.66	7.77
0.66	1946.11	7.73
0.68	1944.57	7.86
0.72	1939.60	8.02
0.74	1936.30	8.22
0.78	1931.84	8.64
0.80	1930.52	8.77
0.84	1926.33	8.47
0.86	1923.55	8.67
0.90	1921.32	8.38
0.92	1913.78	9.33
0.96	1911.17	9.01
0.98	1907.61	9.47
1.02	1899.92	10.31
1.06	1897.06	10.25
1.08	1890.69	11.58
1.10	1887.14	12.32
1.14	1876.68	14.29
1.18	1871.97	14.75
1.20	1849.91	24.57
1.22	1823.47	48.40
1.24	1822.15	39.26

7.8 The chronology of core TCA9b, Tocal Homestead Lagoon

The age estimates for the 2.00 m sediment sequence from Tocal Homestead Lagoon are presented in Table 7.8. Where no direct age determination has been made for a depth in the core, ages have been linearly interpolated from adjacent ages (such ages are denoted by the use of italics).

Table 7.8 The chronology derived for the 2.00 m sequence from core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. The chronology consists of dates compiled from the downcore modelling of excess ^{210}Pb activities, time horizons based on ^{137}Cs activities and lead pollution, palaeomagnetic dating and four AMS ^{14}C determinations. Sediment depths that have not been directly dated have been allocated ages interpolated from adjacent sample depths (these are denoted by the use of italics). Dates derived from radiocarbon determinations have been rounded to the nearest decade.

Depth (m)	Age (years AD)	Error (1 SD) (years AD)	Source
0.00	1998.46	0.00	^{210}Pb
0.02	1997.73	6.69	^{210}Pb
0.04	*1996.35*	*6.54*	
0.06	1994.96	6.39	^{210}Pb
0.08	*1993.35*	*6.43*	
0.10	1991.74	6.47	^{210}Pb
0.12	1990.28	6.61	^{210}Pb
0.14	1988.49	6.80	^{210}Pb
0.16	1987.47	6.91	^{210}Pb
0.18	1986.35	7.04	^{210}Pb
0.20	*1984.79*	*6.96*	
0.22	1983.22	6.87	^{210}Pb
0.24	1982.21	6.98	^{210}Pb
0.26	*1980.25*	*6.86*	
0.28	1978.28	6.73	^{210}Pb
0.30	1976.94	6.87	^{210}Pb
0.32	*1975.57*	*6.76*	
0.34	1974.19	6.65	^{210}Pb
0.36	1973.16	6.76	^{210}Pb
0.38	1971.29	6.96	^{210}Pb
0.40	*1969.79*	*6.79*	
0.42	1968.28	6.61	^{210}Pb
0.44	1966.36	6.80	^{210}Pb, ^{137}Cs (*c.* 1958)
0.46	*1964.31*	*6.79*	
0.48	1962.26	6.77	^{210}Pb
0.50	1960.84	6.92	^{210}Pb
0.52	*1959.16*	*7.06*	
0.54	1957.48	7.19	^{210}Pb
0.56	1955.14	7.47	^{210}Pb
0.58	*1953.36*	*7.50*	

0.60	1951.58	7.52	^{210}Pb
0.62	1949.66	7.77	^{210}Pb
0.64	*1947.89*	*7.75*	
0.66	1946.11	7.73	^{210}Pb
0.68	1944.57	7.86	^{210}Pb
0.70	*1942.09*	*7.94*	
0.72	1939.60	8.02	^{210}Pb
0.74	1936.30	8.22	^{210}Pb
0.76	*1934.07*	*8.43*	
0.78	1931.84	8.64	^{210}Pb
0.80	1930.52	8.77	^{210}Pb
0.82	*1928.43*	*8.62*	
0.84	1926.33	8.47	^{210}Pb, Lead pollution maker (*c.* 1932)
0.86	1923.55	8.67	^{210}Pb
0.88	*1922.44*	*8.53*	
0.90	1921.32	8.38	^{210}Pb
0.92	1913.78	9.33	^{210}Pb
0.94	*1912.48*	*9.17*	
0.96	1911.17	9.01	^{210}Pb
0.98	1907.61	9.47	^{210}Pb
1.00	*1903.77*	*9.89*	
1.02	1899.92	10.31	^{210}Pb
1.04	*1898.49*	*10.28*	
1.06	1897.06	10.25	^{210}Pb
1.08	1890.69	11.58	^{210}Pb
1.10	1887.14	12.32	^{210}Pb
1.12	*1881.91*	*13.31*	
1.14	1876.68	14.29	^{210}Pb
1.16	*1874.33*	*14.52*	
1.18	1871.97	14.75	^{210}Pb
1.20	1849.91	24.57	^{210}Pb
1.22	1823.47	48.40	^{210}Pb
1.24	1822.15	39.26	^{210}Pb
1.26–1.24	*c.* 1820		geochemical and magnetic features
1.28–1.26	1480 (median age)	140	^{14}C, palaeomagnetic feature
1.32–1.30	1290	60	^{14}C
1.94–1.92	-120	70	^{14}C

Chapter Eight

Late Holocene Environmental Change at Tocal

8.1 Introduction

Based solely on the chronology of the lake sediment record, Tocal Homestead Lagoon represents the most detailed and complete record of late Holocene environmental change to date from any site on the Australian continent. The following chapter brings together the results of this research, by merging parallel lines of historical and scientific enquiry. Within the firm chronological framework detailed in the previous chapter, this record will demonstrate how the catchment of Tocal Homestead Lagoon has responded to shifts in human activity and climate, while also providing much-needed information on environmental change in the region during the last 2000 years.

To begin, specific questions on the relationship between human activities, climate and catchment denudation over decadal and centennial timescales will be addressed. In the first instance, this involves some discussion of the role of sediment magnetic mineralogy in recording changes in catchment sediment flux, in the identification of the post-settlement stratigraphy in the lake basin, in the calculation of lake sedimentation rates, and as a proxy record of catchment-wide soil loss. Important research questions raised in the introduction to this thesis may now be addressed. What have been both the immediate and continuing impacts of the arrival of Europeans on catchment soil loss? How rapid has been the catchment's response to shifts in land use? Do any significant time-lags between catchment modification and catchment response exist? How have climatic fluctuations during the last two centuries influenced these processes? To answer these questions with any reliability, a sound understanding of the state of the catchment prior to these impacts is required, as well as a detailed post-settlement record of environmental change.

A broader picture of environmental change in the Tocal region during the late Holocene is then presented, utilising the detailed record of physical, chemical and palynological changes in the sediment sequence. Beginning shortly before 100 BC, the lake sediment record has preserved information on nearly 2000 years of environmental change prior to European settlement in *c.* 1820. This important and rarely examined period of environmental history provides us not only with the immediate background conditions against which the changes wrought by the colonists may be assessed, but, of vital significance, this record spans those millennia when the catchment of Tocal Homestead Lagoon was the home of Aboriginal Australians. Some record of their activities is preserved here, providing the only scientific record to date of their land-use in the region.

While the major focus will be on the environmental conditions before, and immediately after, European settlement, some investigation into environmental change at Tocal up until the present day is also provided, giving the opportunity to examine the long-term consequences of the adoption of European agriculture on the Australian environment. Finally, we may ask how the record from this temperate, near-coastal site compares with the limited information we have on environmental change throughout Australia during the late Holocene?

8.2 Lake sedimentation during the last *c.* 2000 years

8.2.1 Site-specific sedimentation rates

Using the measurements of mineral bulk density (that is, measurements that ignore the contribution of plant organic matter to the sediments) and the chronology derived for core TCA9b, it has been possible to calculate the rate of minerogenic sedimentation at site TCA9 for the last two millennia (Figures 8.1–8.3). This procedure adopts the conservative assumption that the plant organic content of the sediments is entirely the product of vegetative growth in the lake itself, and does not account for any organic material eroded into the lake basin. The procedure also assumes a negligible aeolian sediment contribution to the lake basin. Rates of wind erosion and deposition in the area are low (Woods, 1984; McTainsh and Pitblado, 1987) and this is likely to have been the case throughout the late Holocene. The depositional record in the lagoon is thus likely to reflect changes in the flux of water-borne sediment transported from the surrounding catchment. Recently, several workers have highlighted the potential importance of eroded lake-shoreline material in lake sediment-based studies of sediment budgets (see Lloyd *et al.* [1998], for example). Observations at Tocal suggest shoreline erosion to be minimal however, and this process is considered to be of little importance in contributing sediment to the modern lake basin.

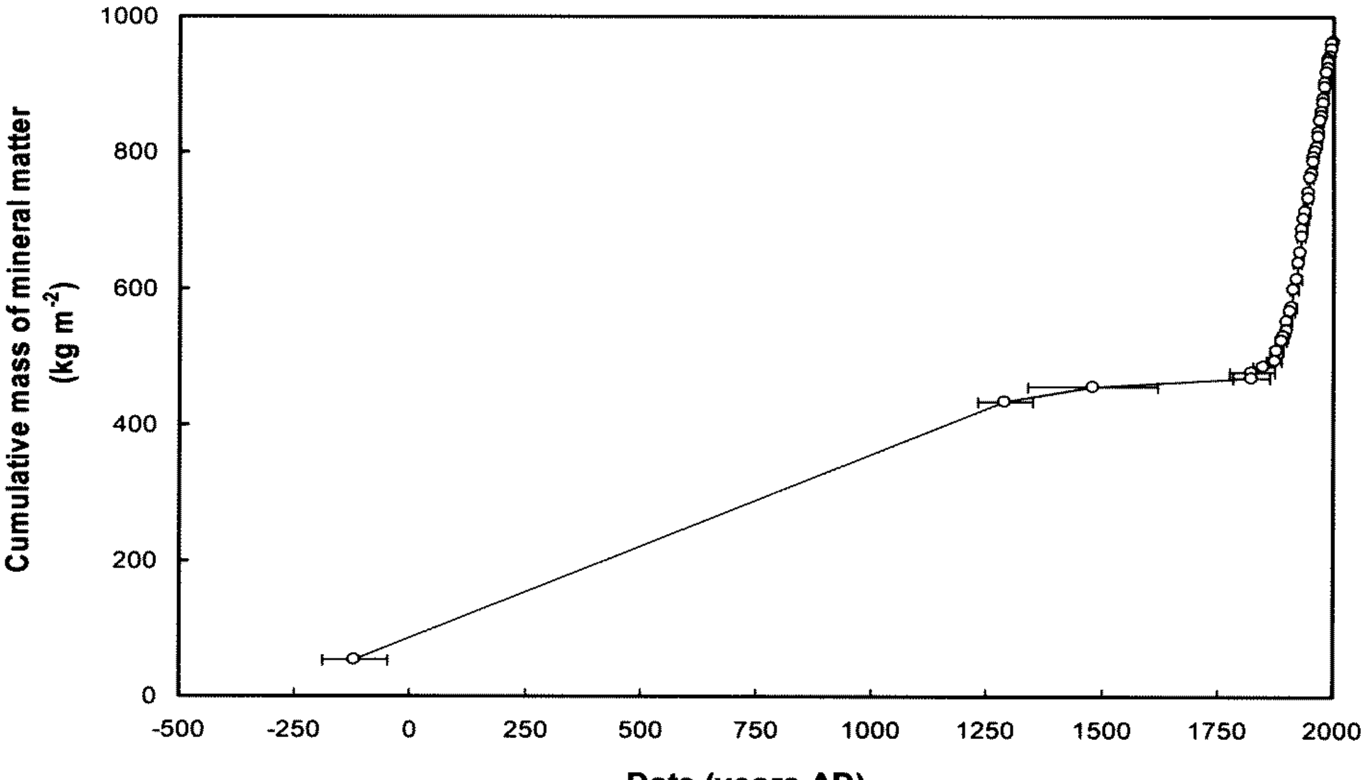


Figure 8.1 The cumulative mass of mineral matter deposited at core site TCA9 for the last *c.* 2000 years, Tocal Homestead Lagoon, Paterson, eastern Australia. Horizontal error bars represent 95% confidence limits on age determinations.

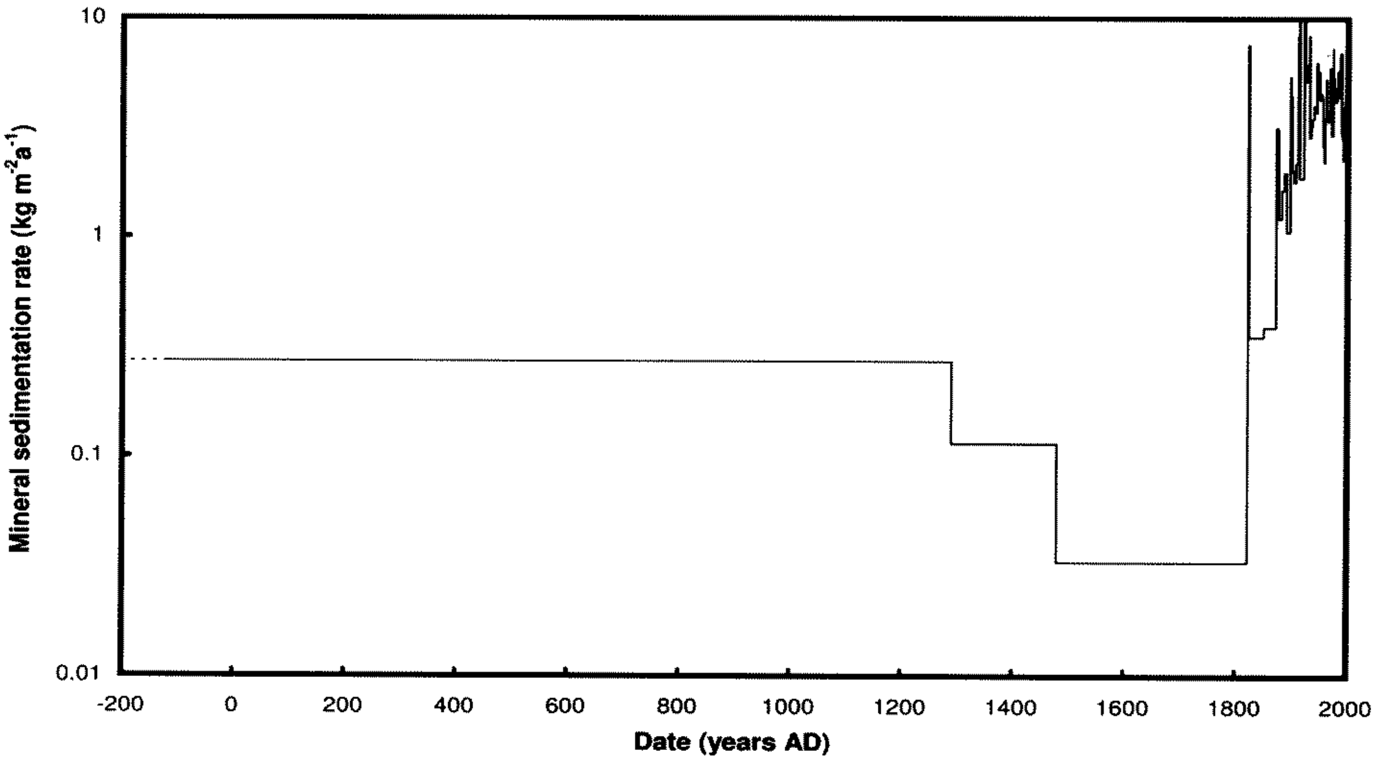


Figure 8.2 Changes in the rate of mineral sedimentation at core site TCA9 for the last *c.* 2000 years, Tocal Homestead Lagoon, Paterson, eastern Australia. For improved clarity, errors associated with these calculations have not been included in the diagram. Note that mineral sedimentation rates are plotted on a logarithmic scale.

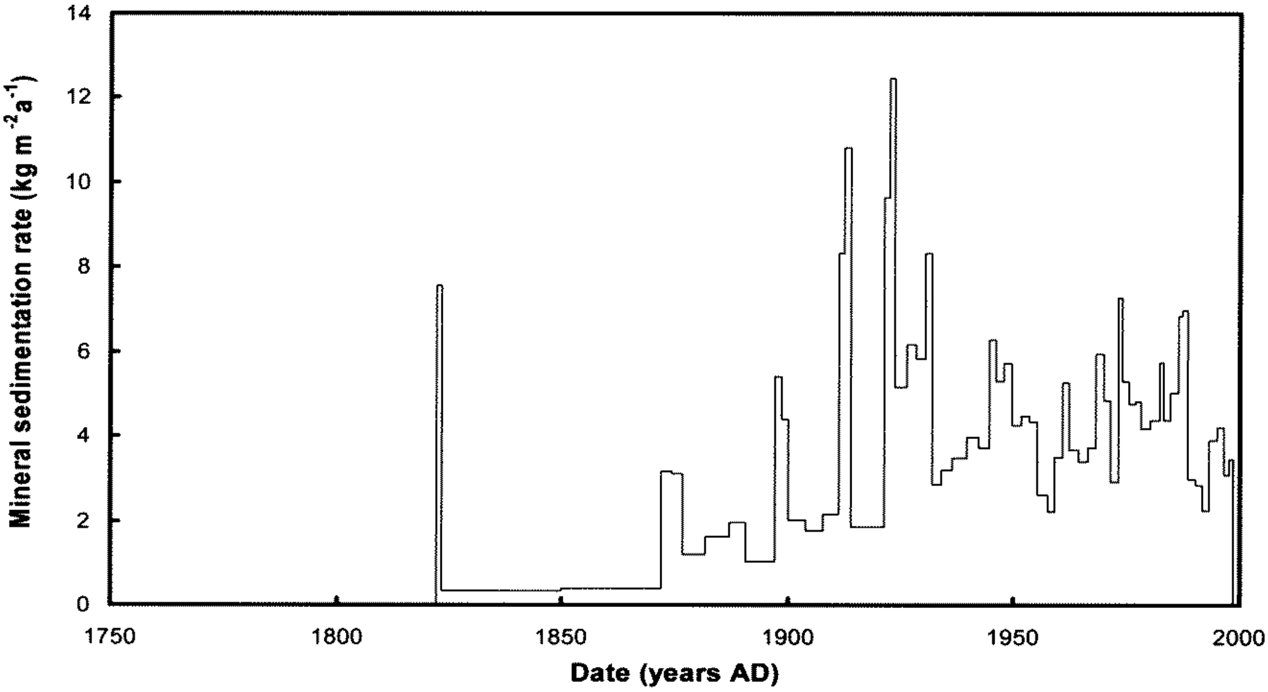


Figure 8.3 Changes in the rate of mineral sedimentation at core site TCA9 for the last *c.* 250 years, Tocal Homestead Lagoon, Paterson, eastern Australia. For improved clarity, errors associated with these calculations have not been included in the diagram.

8.2.2 Lake-wide sedimentation rates

8.2.2.1 Introduction

The pattern of sedimentation in Tocal Homestead Lagoon has remained similar across the lake basin through the last *c.* 2000 years (Figure 6.2). Many of the magnetic features identified in the lake's sediments can be traced across the entire basin; where they cannot, it is often due to the record being distorted at marginal locations (section 7.1.2). The magnetostratigraphy established for the master core may, therefore, be extrapolated across the entire lake basin. Since this is the case, basin-wide sedimentation rates may be determined.

8.2.2.2 Lake sedimentation since European settlement

The magnetic feature labelled 'θ' in Figure 6.2 represents the point in the Tocal sedimentary sequence at which the first significant rise in magnetic susceptibility above background levels is recorded. This feature corresponds with a date of *c.* 1822 AD in the mastercore (1.24 m, TCA9b) and is succeeded by the greatest and most rapid shifts in chemical and magnetic properties recorded in these sediments (e.g., Figure 6.14). Although Europeans first entered the Hunter valley as early as 1790, they did not make their way to the Paterson River until 1801. Convict gangs of cedar loggers may have reached the lower Paterson River by 1818, while the immediate area around Tocal Homestead Lagoon was officially settled by 1822, and partially cleared from this date onwards. Initial environmental disturbance in the catchment of Tocal Homestead Lagoon is most likely to have occurred, therefore, between 1818 and 1822 (section 4.4.1).

The date of feature 'θ' closely corresponds with the arrival of the first Europeans at Tocal, as determined from the historical record. The overlying sediments (above the 1.26 m horizon) most probably represent the first record of European disturbance in the catchment. As this feature may be traced across the majority of the lake basin, the depth and volume of sediment deposited in the basin since European settlement may be determined. Utilising the master core's detailed record of changes in the organic component and bulk density of the sediments, an estimate of the volume of minerogenic sediment laid down in the lake since the early 1820s can be made.

The total volume of sediment transported into the lake since *c.* 1822 is 93 000 m^3. Using the mineral bulk density of the sediments (section 6.2.5) overlying the 'θ' horizon, the mass of minerogenic material eroded into Tocal Homestead Lagoon since European settlement may be estimated as 38 000 t. Some qualifications must be made in presenting this figure. By subtracting the volume of material lost on ignition at 430°C (section 6.2.4) from the total volume of sediment deposited, the conservative assumption has been made that the plant organic content of these sediment is the product of the biological activity in lake itself, rather than material eroded from the surrounding catchment. Support for this assumption comes from the visual record of core TCA9b, in which the majority of visible organic material is made up of aquatic plant roots in growth position (section 6.2.1). Secondly, it is assumed that the inorganic material in the lake is composed entirely of material eroded from the catchment. As detailed in section 8.2.1, wind deposition and erosion in the region is low (Woods, 1984; McTainsh and Pitblado, 1987), therefore a significant atmospherically-derived contribution to the sediments is unlikely. The contribution of sediment to the lake as a result of overflow from Webber's Creek or the Paterson River is also thought to be negligible. Careful visual examination of the mastercore found no evidence of flood-borne sediments, while a comparison of the magnetic signatures of suspended sediment samples collected from the Paterson River by Doyle (1999) provides even clearer proof that there is no significant sediment component entering the lake from this source (Figure 8.4). Finally, it is assumed that the lake basin is essential enclosed and that no sediment has been lost. Although this has been the case for the last 2000 years, the lake's modern outlet has meant that the volumes of sediment calculated for the most recent decades may be underestimates.

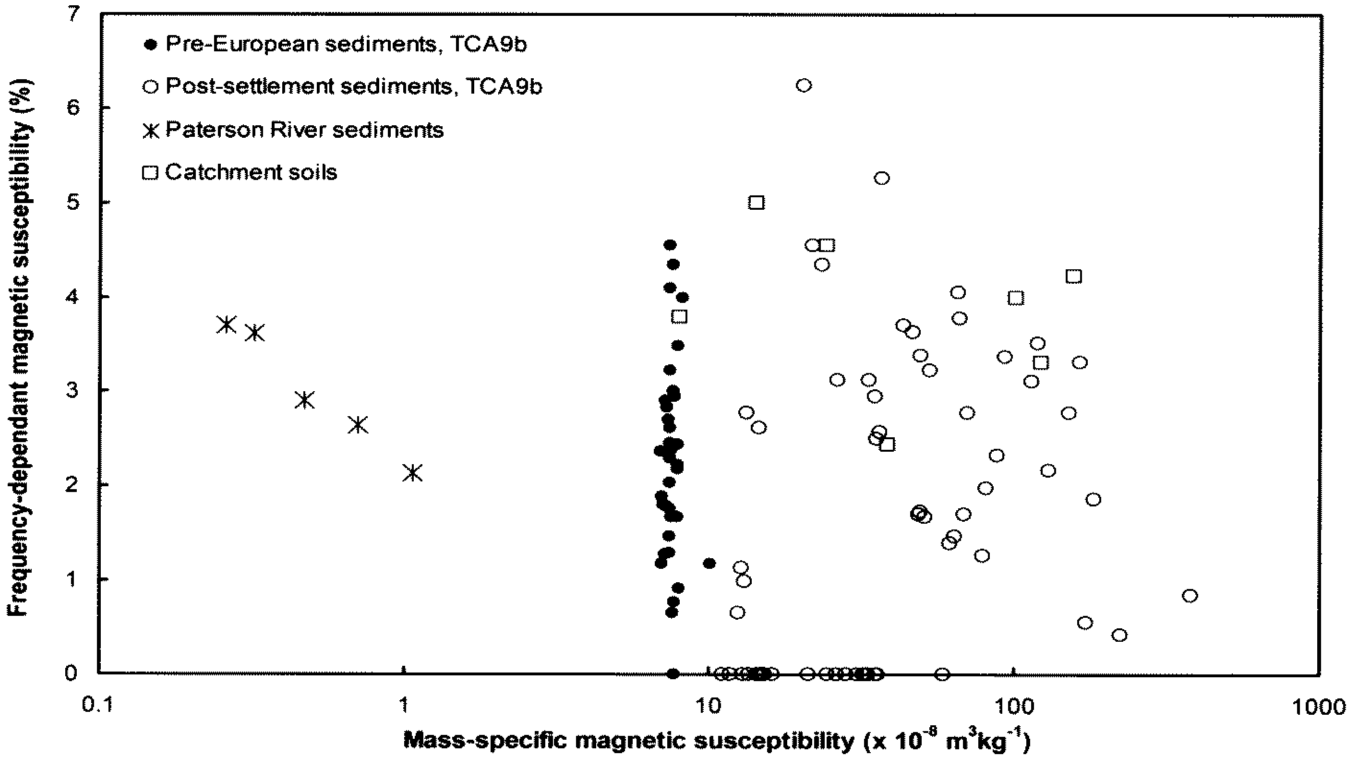


Figure 8.4 Magnetic properties (mass-specific and frequency-dependant magnetic susceptibility) of core TCA9b, Tocal Homestead Lagoon, catchment soil samples (Cook, 1998) and suspended sediment samples from the Paterson River upstream of Tocal (Doyle, 1999). The provenance of the lake sediments is most similar to the soil samples from the catchment, and none of the lake sediments displays any similarity with those sediments from the Paterson River, which are characterised by low magnetic susceptibilities.

8.2.2.3 The spatial pattern of sediment accumulation since c. *1822 AD*

The spatial pattern of sedimentation in the lake since European settlement suggests that processes of overland flow have dominated the delivery of eroded materials from the surrounding catchment to the basin. Sedimentation in the lake has been concentrated along the basin's central axis and southeast corner (Figure 8.5), primarily near the deep-water locations of A8 and A9 (Figure 3.5). There is little indication that fluvial input into the basin has provided any significant contribution to basin sedimentation. There is no evidence of deltaic accumulations near the mouth of the inlet stream, and no suggestion of any decrease in the volume of sediment deposited away from the southwest limb of the lake.

An important insight into Tocal's history may be gained from this chart; this is that the greatest volume of sediment has originated from the east and southeast slopes of the lake (Figure 8.5). Although the drainage lines that enter the basin along these shores today are essentially inactive (Figure 8.6), evidence from historical aerial photographs indicates that this has not always been the case (Figure 8.7). Nevertheless, concentrated flow to the lake appears to have had little effect on the pattern of sedimentation. Importantly, this pattern has provided some indication of where in the catchment agricultural activities may have been concentrated in the past. Although the historical record developed for Tocal is quite comprehensive, one shortcoming has been the lack of information on where cultivation and grazing took place at the site.

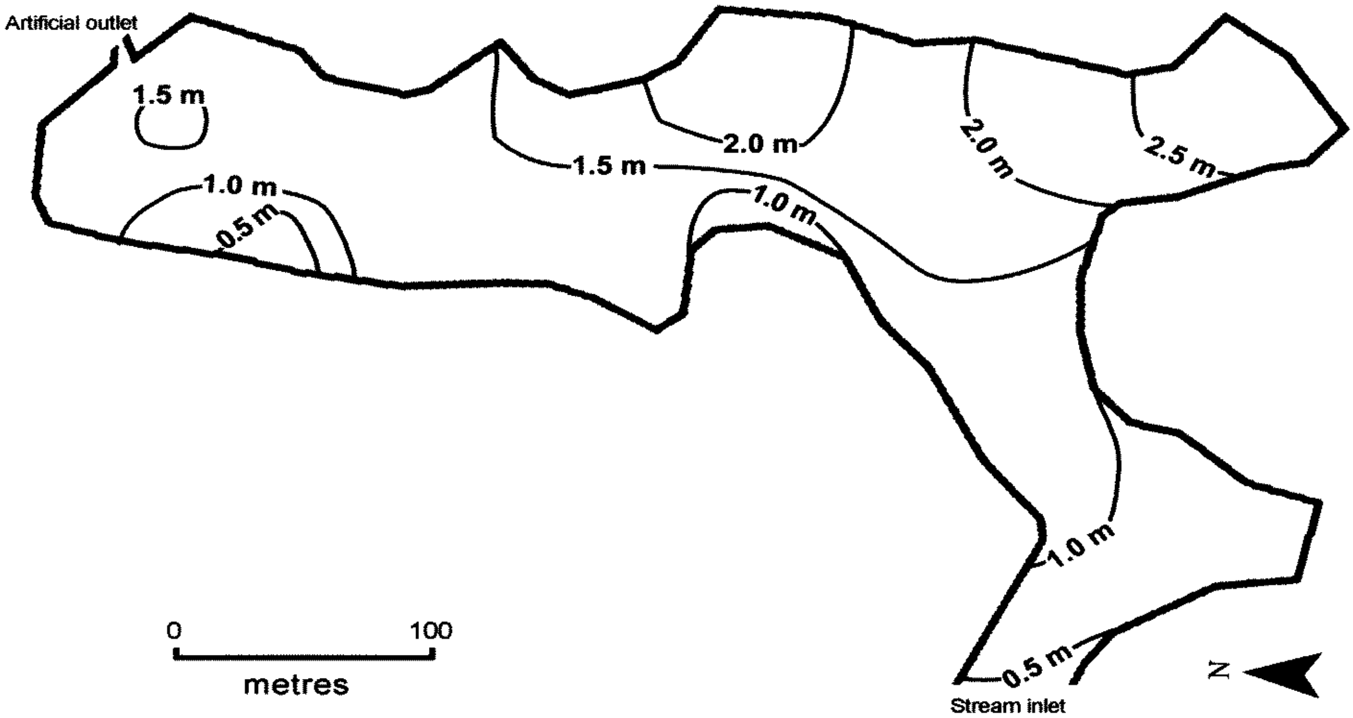


Figure 8.5 The depth of sediment deposited in Tocal Homestead Lagoon, Paterson, eastern Australia since the arrivals of Europeans at the site in *c.* 1822.

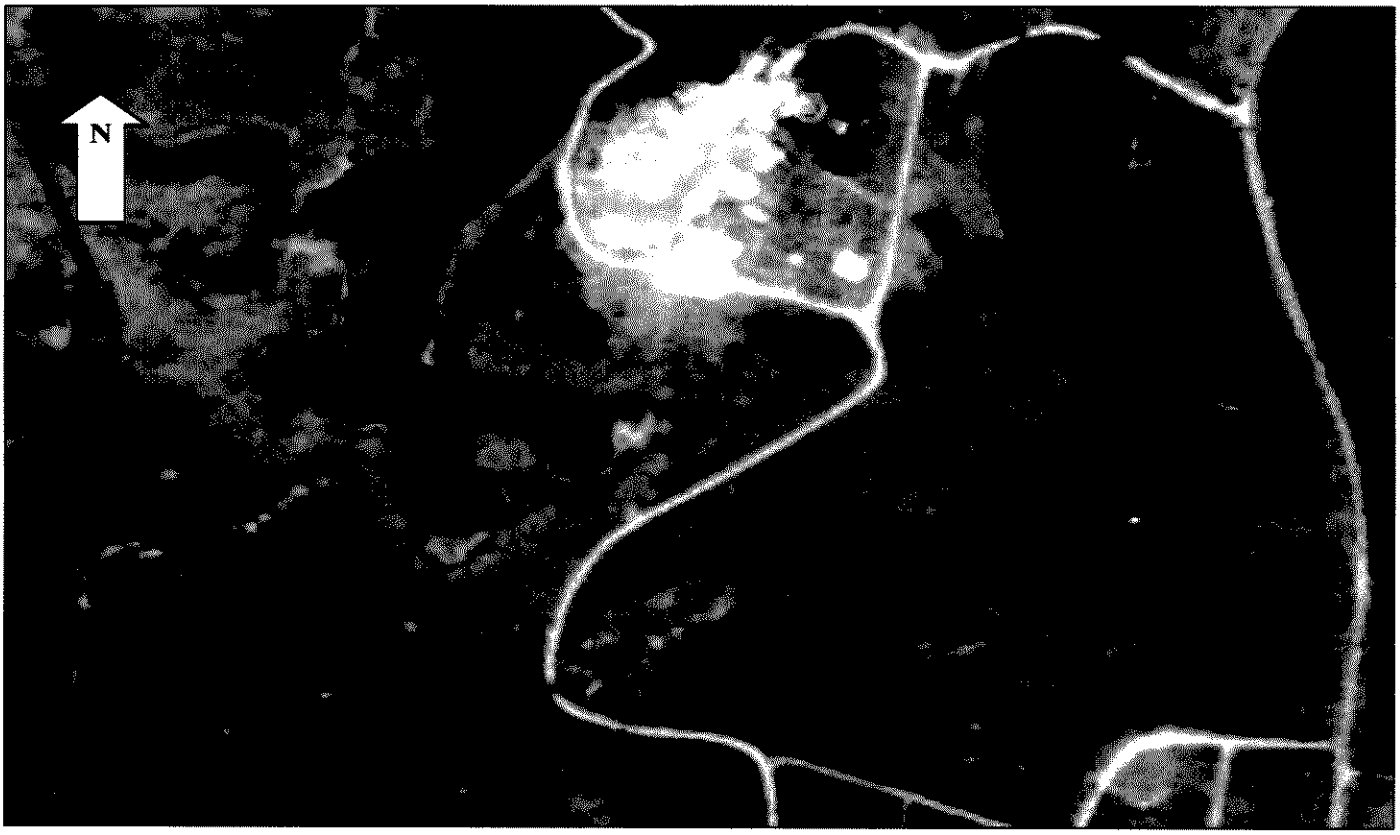


Figure 8.6 Vertical aerial photograph of Tocal Homestead Lagoon, Paterson, eastern Australia in 1996.

Figure 8.7 Vertical aerial photograph of Tocal Homestead Lagoon, Paterson, eastern Australia in 1967.

8.2.3 Mean annual catchment-wide erosion rates at Tocal

Numerous studies have shown how lake sedimentation data may be used to obtain estimates of catchment-wide denudation rates (Davies, 1976; Dearing, 1983, 1992; Dearing *et al.*, 1981, 1990; Foster and Walling, 1994; Foster *et al.*, 1986, 1990, 1996; Page and Trustrum, 1997; Gale and Haworth, 2005). Theoretically this is simple enough: the mass of mineral sediment deposited in an enclosed lake basin during a given period of time may be converted to the mass of material stripped from the surrounding catchment for that same period, and so rates of soil erosion averaged for the area of the catchment may be made. In practice, however, this procedure is founded on a range of assumptions that may rarely be met in any individual system. In addition to the assumptions made about the system in section 8.2.2.2, the presumption is made that material eroded from the catchment has been transported near-instantaneously into the lake basin, such that changes in lake sedimentation rates are near-contemporaneous with shifts in catchment erosion processes. The Tocal Homestead Lagoon catchment is small and is characterised by short and relatively steep slopes. It has no major topographical depressions other than the lake basin itself. Most of the catchment slopes are directly connected to the lake basin, while those that are not (in the catchment's southwest corner) drain into the lake via an efficient stream network (Figure 2.1). In this situation, the assumption made regarding direct sediment delivery is defensible. However, any estimates of catchment denudation made based on lake sedimentation will inevitably underestimate true rates of removal from slopes, for a small, and realistically undeterminable, volume of eroded material will be stored on the lower gradient segments of catchment slopes at least temporarily.

Following this procedure, a minimum catchment-wide denudation rate of 40 300 t km^{-2} may be calculated. For the 176 year period since European settlement, this translates to a mean annual rate of mineral soil loss of at least 228 t $km^{-2}a^{-1}$. Examination of the long-term mean sedimentation rates recorded in the mastercore shows that these increased from 0.21 in the two millennia prior to European settlement to 2.82 kg $m^{-2}a^{-1}$ (a factor of 13) in the 176 years since. This may be used to provide an estimate of the mean rate of mineral soil loss for the *c.* 2000 years before European settlement: at least 18 t $km^{-2}a^{-1}$.

8.2.4 The long-term impact of European settlement on erosion in eastern Australia

To understand what happened to soils and landscapes following the arrival of Europeans and modern agriculture in Australia, continuous environmental records that not only span the period of European settlement, but also pre-date it are required. Apart from Tocal Homestead Lagoon, there are only three well-dated sites on the Australian continent capable of providing reliable data for testing the long-term impact of early European agriculture on soil erosion. These are Little Llangothlin Lagoon in northeast New South Wales (Gale and Haworth, 2005), Redhead Lagoon in central eastern New South Wales (Gale, 2003) and Umberumberka Creek in far western New South Wales (Wasson and Galloway, 1986) (section 3.6.2). As these three sites span a range of geographical environments in eastern Australia, a comparison of their results may provide a representative picture of how sites in the eastern half of the Australian continent have responded to European land-use.

As detailed in section 3.6.2, the records from these sites show that the mean rates of sediment flux from catchments in eastern Australia have increased by between seven and 66 times since European settlement. It is possible that the greatest impacts were experienced in drier and poorly-vegetated areas such as those of far western New South Wales (Umberumberka Creek), while impacts were least in the well-vegetated humid temperate environs of the New England Tablelands (Little Llangothlin Lagoon), but the data are not yet sufficient to do more than speculate on this. A greater range of sites with comparable records will be needed to better understand the broader role that climate has played in controlling sediment yields across Australia.

8.2.5 Comparison of the Tocal catchment sediment yield with other sites

To gain some insight into the magnitude of accelerated soil loss at Tocal, the mean annual rate estimated for the period since European settlement may be evaluated against sediment yield data from other sites in Australia. A comparison may be made first by comparing the Tocal rate with those obtained from catchments throughout New South Wales. The majority of catchment sediment yield data are collected by stream gauging, while erosion plot and sediment tracer/budget studies (such as those using ^{137}Cs methods), and reservoir-infill reconstructions are much less common (section 3.5.4). Only two sites in Australia, Little Llangothlin Lagoon (Gale and Haworth, 2005) and Umberumberka Creek (Wasson and Galloway, 1985) have produced long-term catchment-wide rates of erosion that date from the onset of European settlement. Direct comparison, therefore, between most previous research and the Tocal data is not possible. Nonetheless, it is still important to test these findings against the published record. In doing so, the lake-sediment based approach adopted at Tocal may be checked for efficiency against the more usually employed sediment yield techniques.

A first-order comparison may be made using the compilation of sediment yield data from Olive and Walker (1982). To place the post-settlement catchment sediment yield from Tocal in perspective, it lies at the top end of yields from catchments of similar sizes from throughout New South Wales. While Gale and Haworth (2005) compared their lake sediment estimates of catchment denudation from Little Llangothlin Lagoon with those from upland southeast Australian sites collated by Wasson (1994), a more appropriate comparison for the rates from Tocal is with the extensive body of soil erosion data collected from the Hunter valley region during the last several decades (Figure 8.8).

As commonly reported in the literature (Walling, 1983; 1984; Walling and Webb, 1996, and references therein), an inverse relationship between sediment yield and catchment area is exhibited over the range of Hunter valley sites examined. Although the Hunter valley exhibits some climatic variation—particularly differences between the coastal and inland parts of the region, and those high-elevation regions around Barrington Tops—in general, climate, land-use and soil types throughout the valley are broadly similar to those at Tocal. Importantly, many agricultural sites in the region have experienced comparable histories of land-use, with most of the lands 'improved' by the late 1820s. Such sites have been obvious targets of sediment yield studies, such as Loughran *et al.*'s (1992) investigation of hillslope erosion in Hunter valley vineyards. The mean rate of soil loss for the last 176 years at Tocal is higher than any rate recorded in catchments of similar size in the region. However, a comparable rate of 215 t $km^{-2}a^{-1}$ was reported by Loughran *et al.*

(1992) at Maluna Creek, Pokolbin for slopes under intensive agriculture (viticulture). In contrast, the closest site to Tocal with catchment sediment yield data, a 1.2 km^2 catchment near Wirragulla in the Williams River catchment (25 km east of Tocal) recorded a mean catchment-wide soil loss of just 19 t $km^{-2}a^{-1}$, determined using ^{137}Cs methods (Krause *et al.*, 2003a).

Of all these sites, perhaps the closest comparison may be made with the farm-reservoir based reconstructions of Erskine *et al.* (2002) from twelve <0.1 km^2 catchments in the southern Hunter valley and northern Sydney region. Like Tocal, these sites are essentially sealed systems, where over long timescales the total flux of soil from the catchment is retained, providing an accurate estimate of catchment-wide soil loss. Erskine *et al.* (2002) provided three mean sediment-yield rates from these sites; for catchments under cultivation (710 t $km^{-2}a^{-1}$), under grazing (320 t $km^{-2}a^{-1}$) and under forest (310 t $km^{-2}a^{-1}$). These results are comparable with those from Tocal. All three rates may grossly overestimate catchment sediment yield, however, as they appear to include the organic fraction of the dam sediments, which in productive lakes in the region may be as high as 50% by mass (Williams, 2005).

Is the relatively high catchment-wide soil loss rate from Tocal the product of exceptionally severe land-use or its geomorphic setting? The answer is probably no to both questions. Although the site has a long history of European agriculture, it is not singular in the variety and intensity of activities that have characterised it, particularly when viewed from the perspective of catchment degradation. The catchment has no major relief, nor exceptionally steep slopes. The most probable explanation is related to the efficiency of sediment delivery into Tocal Homestead Lagoon, in contrast with those larger fluvial basins that rely on suspended sediment yield data for estimates of basin erosion. Such sites will inevitably underestimate erosion rates due to the inefficiency of sediment delivery through the basin (Walling, 1983). However, sediment yields from hillslopes do not appear to be greater than those from river basins in the Hunter valley, suggesting that increasing basin-size, and increasing delivery retardation may not be the only confounding phenomena.

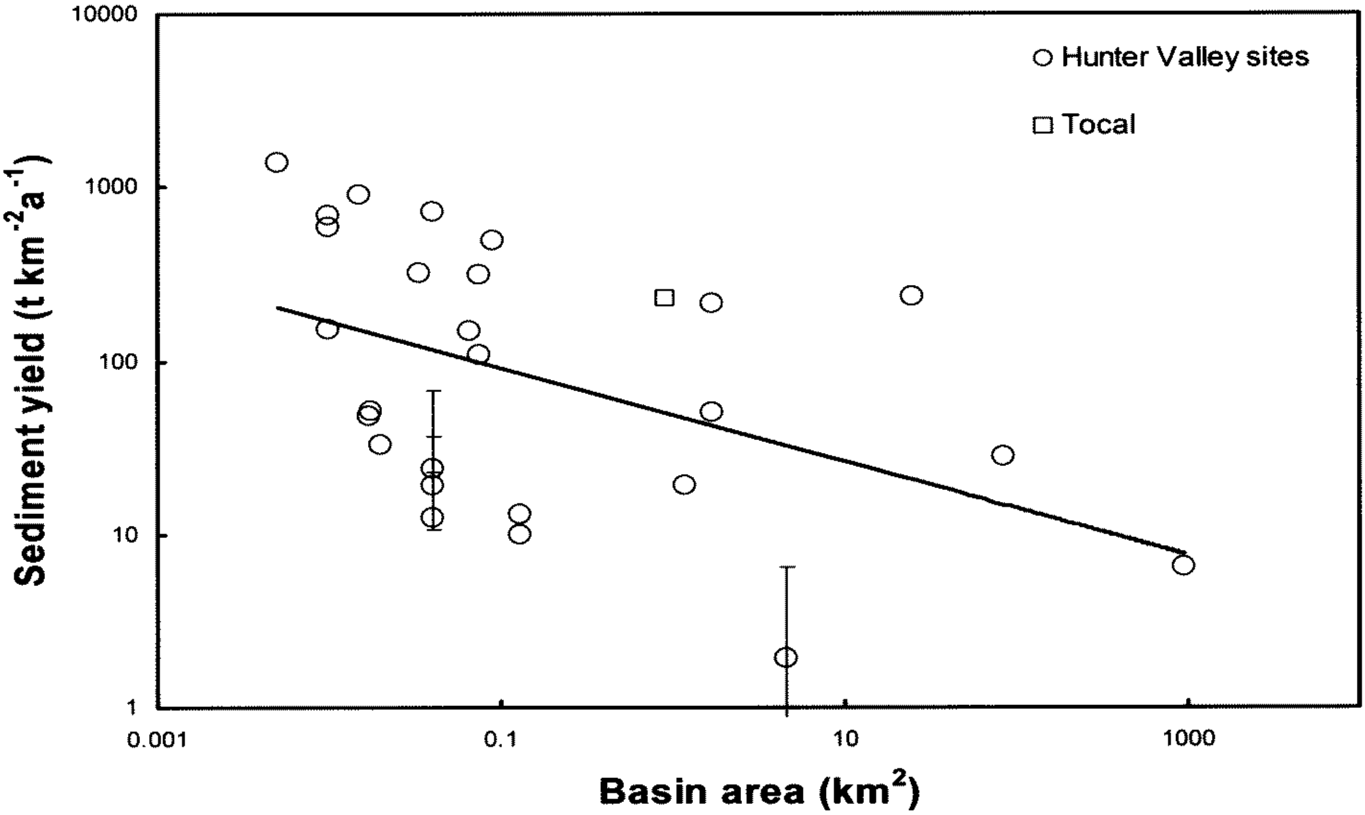


Figure 8.8 Basin areas and corresponding sediment yields determined for catchments in the Hunter valley region. Data from: Loughran (1977), Geary (1981), Campbell *et al.* (1986), Loughran *et al.* (1986, 1988, 1990, 1992, 1993), Morris and Loughran (1994), Elliott *et al.* (1997), Erskine *et al.* (2002), Krause *et al.* (2003a, 2003b). Vertical error bars span the range of sediment yields recorded at some sites.

Those sites in the Hunter valley monitored over the short-term using hillslope plots (Loughran *et al.*, 1993) and ^{137}Cs methods (Krause *et al.*, 2003a) have generated lower than average results, and may also have underestimate catchment sediment yields. As the stratigraphy from Erskine *et al.*'s (2002) reservoir sediments suggest, erosional processes in this region may be entirely episodic, therefore short-term monitoring may miss major episodes of erosion, generating misleadingly low results. Sites monitored year-round in the Hunter valley show that as much as 99% of sediment yield is transported from slopes in only half of the year (Loughran, 1977; Geary, 1981a [cited by Loughran, 1984]). Perhaps the length of record (and, therefore, experimental approach) is a more important factor in explaining variations in sediment yields for basins of similar areas than whether an entire basin or a single slope is monitored?

The mean post-settlement catchment-wide soil loss rate for Tocal is, however, limited in providing detailed information on how the catchment has responded to changing land-use and climatic conditions. It is obvious from Figure 8.3 that there have been enormous variations in sediment flux since the arrival of Webber at Tocal in the early 1820s; the mean catchment-wide soil loss rate of 238 t $km^{-2}a^{-1}$ effectively masks this record.

8.3 A high-resolution record of soil erosion since *c.* 1820 AD

8.3.1 Determining rates of catchment sediment yield at Tocal

Using the high-resolution chronology from the master core, and the site-specific sedimentation rate established at core location TCA9, a detailed record of changes in sediment flux from the Tocal catchment since European settlement in the early 1820s can be developed. The basin-wide magnetostratigraphy established at Tocal demonstrates that the pattern of sedimentation observed in the master core record is representative of the lake basin, rather than simply a record of localised lacustrine processes. Temporal changes in catchment processes, therefore, recorded at this core location may be taken as representing the entire catchment's erosional history. Since this is the case, the chronology and mineral bulk density record from the master core may be used to determine the cumulative mass of mineral matter deposited in Tocal Homestead Lagoon since *c.* 1822 (Figure 8.9). Adopting the assumptions made in section 8.2.2.2, these data may be expressed as catchment-wide rates of minerogenic erosion, providing estimates of erosion rates from the lake's catchment from *c.* 1820 to present (Figure 8.10).

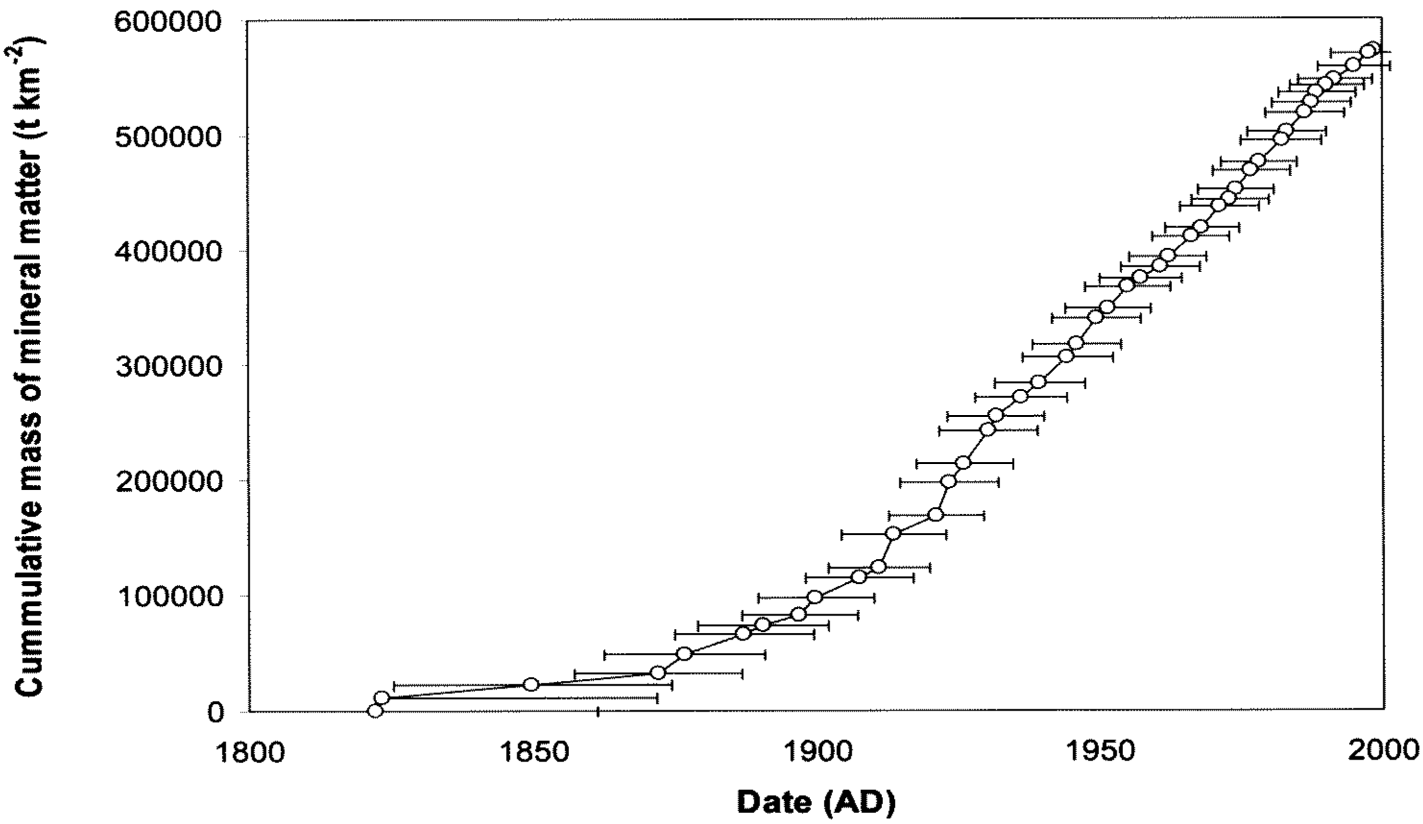


Figure 8.9 The cumulative mass of mineral matter deposited in Tocal Homestead Lagoon, Paterson, eastern Australia for the period *c.* 1820 to 1998. Horizontal error bars represent 95% confidence limits on ^{210}Pb ages. However, the large lower error bar on the lowest ^{210}Pb age has been removed, as the age of the sediments from this depth has been tightly constrained using palaeomagnetic and ^{14}C dating, and sedimentological evidence (Chapter 7), showing the central ^{210}Pb age to be accurate.

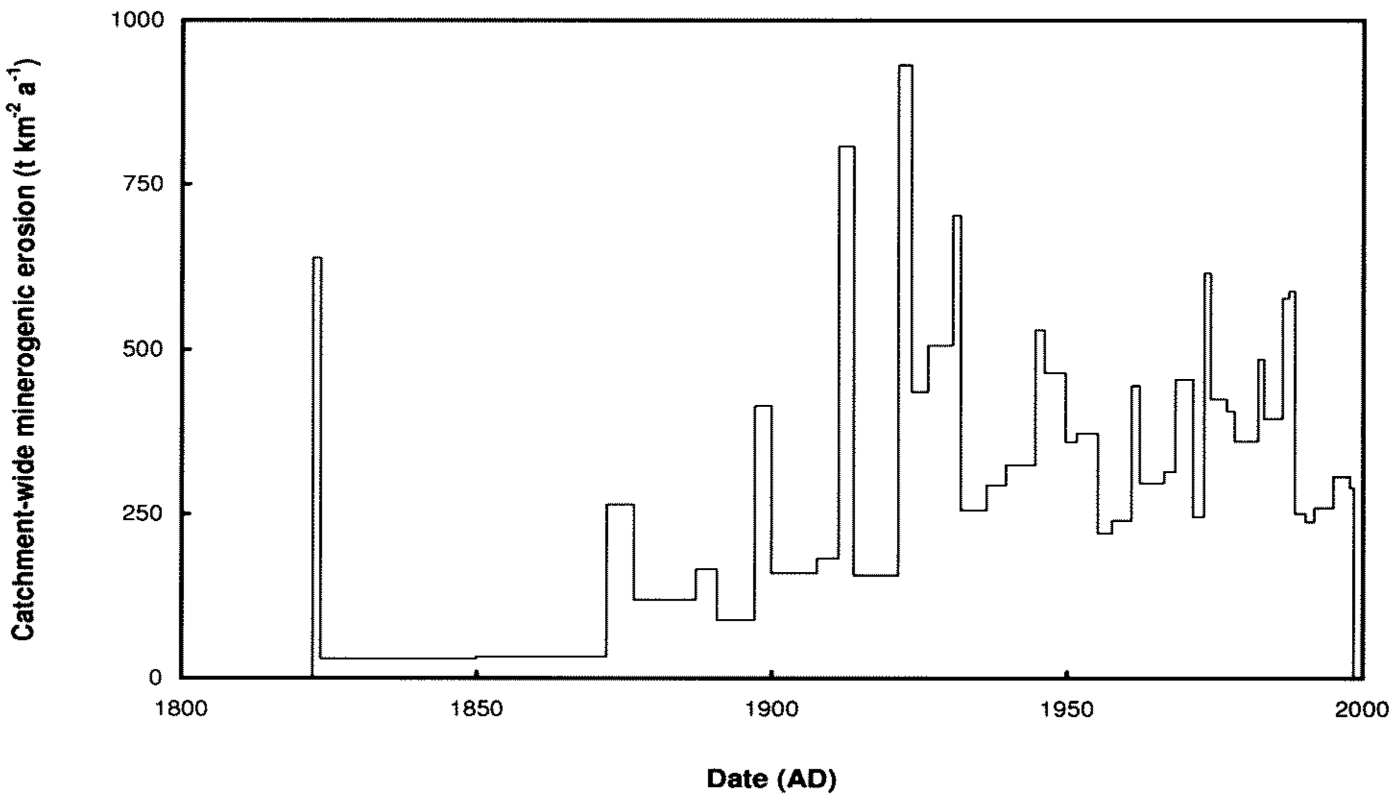


Figure 8.10 Catchment-wide rates of minerogenic erosion from the catchment of Tocal Homestead Lagoon, Paterson, eastern Australia for the period *c.* 1820 to 1998.

8.3.2 Pre-European catchment-wide rates of erosion

So far, the only information we have on the magnitude of environmental change that accompanied the arrival of Europeans at Tocal comes from changes in site-specific sedimentation rates (Figure 8.2). These show that the rate of lake sedimentation at core location TCA9 increased immediately upon the arrival of the first settlers by over 240 times. To gain an understating of the catchment processes that were operating in the millennia before this and to establish the 'baseline' conditions before the 1820s AD, a simple comparison of the post-settlement catchment-wide rates can be made with the site-specific sedimentation record for the last 2000 years (section 8.2.1). On a proportional basis alone, a crude estimate of catchment sediment yields may be made for the late Holocene before European settlement (Figure 8.11). Again, the assumption is made that the sedimentation pattern exhibited at the master core location is representative of sedimentation basin-wide. For the post-settlement period at Tocal, this is justifiable, with the basin-wide magnetostratigraphy confirming this. However, the processes of erosion and lake deposition in the catchment prior to *c.* 1820 are unlikely to have been identical to those of the last two centuries, and this assumption is more difficult to accept. Even so, the errors arising from this are probably minor and relatively unimportant as long as the relative magnitudes of pre- and post-settlement rates are maintained.

Catchment-wide sediment yields of minerogenic material have thus been determined for three periods before European settlement: *c.* 100 BC to *c.* 1300 AD (23 t $km^{-2}a^{-1}$), *c.* 1300 AD to *c.* 1500 AD (10 t $km^{-2}a^{-1}$) and *c.* 1500 AD to *c.* 1820 AD (3 t km^{-2} a^{-1}) (Figure 8.11).

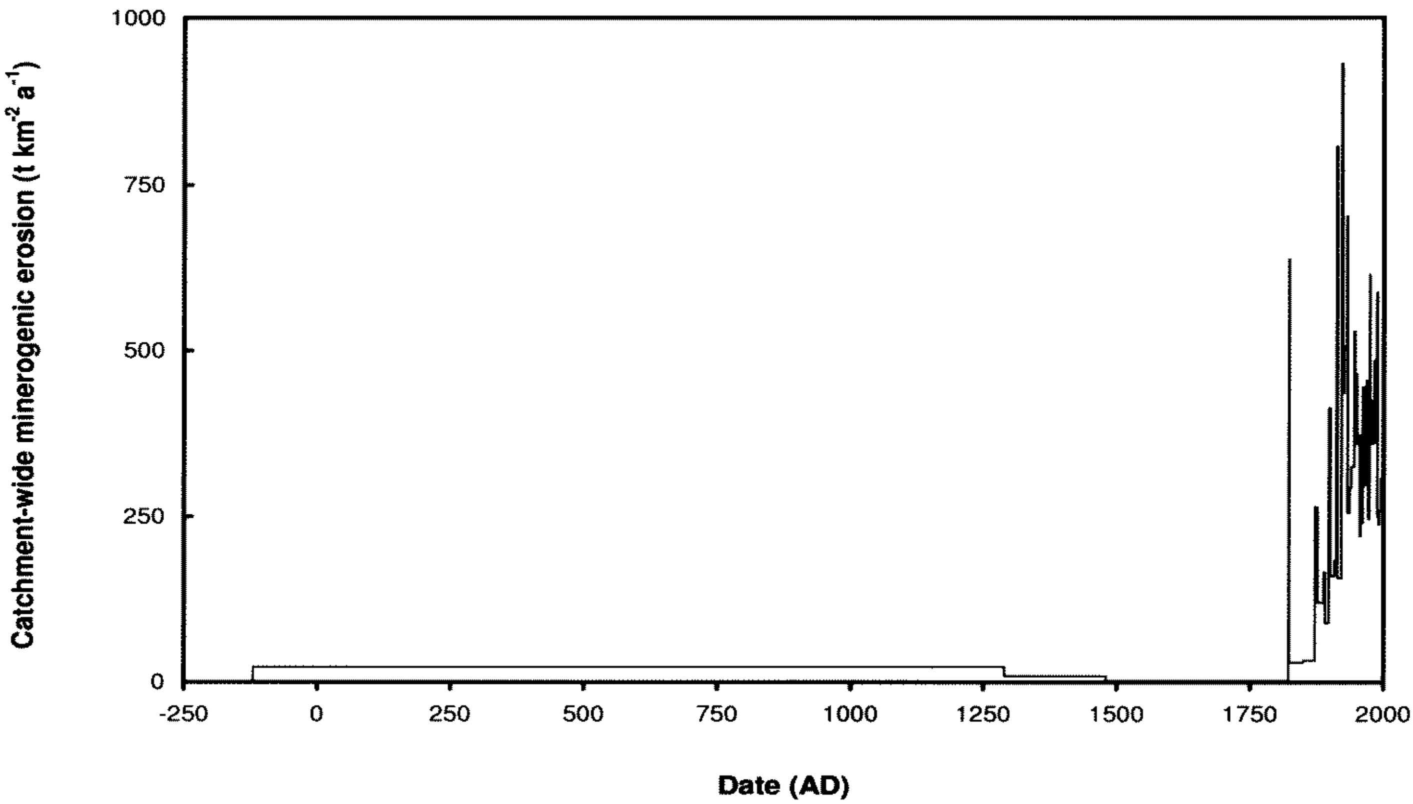


Figure 8.11 Catchment-wide rates of minerogenic soil loss from the catchment of Tocal Homestead Lagoon, Paterson, eastern Australia for the period *c.* 100 BC to 1998 AD.

8.3.3 Estimates of pre-European soil loss from eastern Australia

How do the estimates of catchment denudation from Tocal compare with the limited available data we have for the period before European settlement? Are they typical of an undisturbed system? One estimate of pre-European sediment yields from southeast Australia's coastal fringe by Harris (1995, cited by Wasson *et al.* [1996]) is 3.9 t $km^{-2}a^{-1}$. Being based on modern measurements of 'undisturbed' catchments, this value is most likely an overestimate, as Wasson *et al.* (1996) noted for Harris' estimate of sediment yield from the Murray-Darling basin of eastern Australia. Other similar estimates of pre-disturbance erosion rates based on modern measurements, usually obtained from averaging the lowest recorded yields in a region, are detailed in section 2.5.4.

Only one other study from Australia has provided comparable data on pre-European rates of catchment denudation: Little Llangothlin Lagoon (Gale and Haworth, 2005). Gale and Haworth (2005) calculated the rate of soil loss from the catchment of Little Llangothlin

Lagoon, northeast New South Wales, Australia, to be 25 t $km^{-2}a^{-1}$ for the three decades before European settlement in the basin. This is essentially the same as the mean rate of soil loss at Tocal for the 1500 year period before 14th century AD. It should be noted, however, that Gale and Haworth (2005) have argued that this rate itself is inflated by human impact. Interestingly, catchment-wide soil loss at Tocal decreases after this time, reaching a minimum rate of just 3 t $km^{-2}a^{-1}$ in the period from *c.* 1500 AD to European contact. Does this long-term decreasing trend lie within the amplitude of natural variations in sediment flux, or is it a record of changing processes in the catchment centuries before European settlement? In conjunction with further evidence from the Tocal sediments, these questions will be addressed in section 8.4.

8.3.4 The impact of European settlement on soil loss

With a solid understanding of antecedent erosion conditions at Tocal, one of the main questions of this research raised in Chapter 1 may be addressed: what was the impact of European settlement at Tocal on catchment soil loss? Immediately following the arrival of European settlers *c.* 1822, catchment-wide minerogenic erosion rates increased enormously, from 3 t $km^{-2}a^{-1}$ to 638 t $km^{-2}a^{-1}$ (Figure 8.11), an increase of over 210 times. Not only was the catchment's response to deforestation and European agriculture (section 4.4) severe, but it was swift; there is no lag period between initial catchment modification and catchment response. It appears that this sharp increase in erosion took place very shortly after European settlement. This corroborates the story obtained from other high-resolution records of soil erosion in Australia (Gale, 2003) and controverts the theory championed by some researchers that the immediate impacts of European settlement on the Australian environment were relatively minor (section 2.6.2).

Several factors, acting individually or in combination, have been suggested by Gale (2003) to explain the dramatic response to settlement. The trampling of soft, uncompacted pre-contact soils by livestock would have led to lowered infiltration capacities, increased runoff and enhanced sediment yield. The discontinuous nature of the native grass cover would have incompletely bound the soil and exposed much of its surface to erosion. The removal and modification of the pre-contact vegetation cover by grazing, burning and clearance would have exposed and disturbed the soil, making it more susceptible to erosion. In the case of Tocal, we have evidence that at least one of these factors may have been significant. Soil profiles preserved beneath the original (*c.* 1822) homestead are very friable, poorly structured and have weak aggregate stability (section 6.9), all features characteristic of high erodibility.

8.3.5 Soil formation rates and sediment supply in the last 2000 years

With catchment-averaged rates of erosion reaching levels as high as 931 t $km^{-2}a^{-1}$ at Tocal in the past, what effect has this had on the inevitably limited supply and quality of soil? Only a small body of information exists on soil formation rates in Australia, and most of this concerns soil formation rates on unconsolidated parent material (Walker and Coventry, 1976; Costin, 1980; Thompson, 1981; Walker, 1981; Edwards, 1988; Edwards and Zierholz, 2000). Although limited in their geological scope, these data are of some relevance to Tocal as some of the soil types found in the catchment have formed on Quaternary alluvium. Beckham and Coventry (1987) and Edwards (1988) have documented the sluggish rates of soil formation in Australia, suggesting that they are so low as to be practically immeasurable, especially in more arid areas. Maximum rates of formation are considered to be around 40 t $km^{-2}a^{-1}$ (Gale and Haworth, 2005). Rates of formation on consolidated parent material, even in the weathering-intense environments of tropical northeast Australia, are thought to be up to three orders of magnitude slower (Pillans, 1997). Assuming that rates of soil formation at Tocal lie close to the maximum values determined for Australian soils on alluvial parent material, then rates of catchment denudation must have been lower than formation rates for the two millennia before European settlement. For the centuries immediately before their arrival, soil formation rates may have been as much as an order of magnitude greater than the rate at which soil was being removed from the catchment slopes.

During the early 1820s, catchment denudation was more than 40 times the assumed rate of soil formation. Any 'stockpile' of soil resulting from the at least 2000 year period of lower erosion would have been near-depleted in the following 180 years. If the initial peak erosion rate of 638 t $km^{-2}a^{-1}$ continued for only the first five years of European settlement at Tocal, over 200 year's worth of soil material would have been stripped from the catchment's slopes. As agricultural activity was confined to only portions of the Tocal property, the catchment-wide erosion rate for the early 1820s may even underestimate the real rates of erosion, and the impact of these activities on the immediate slopes of the lake. But by how much? With accelerated erosion coming from only half of the total catchment of Tocal immediately following European settlement (a conservative overestimate), localised rates of erosion may have been as high as 1200 t $km^{-2}a^{-1}$ from these slopes, rapidly stripping any erodible material into the lake. Such severe denudation would translate to the loss of thousands rather than hundreds of year's worth of soil formed from alluvial parent material.

8.3.6 Climate and erosion since European settlement

An important component of this research has been to understand the role that climatic perturbations have played in controlling catchment erosion, both before and after European settlement. However, the role of climatic conditions through this period, and before Webber's arrival in the catchment cannot be considered independently of land use. Over long time spans, a range of positive and negative feed-back loops exists between the two broad processes, with climate inevitably influencing human activity, while the role of land use in modifying climatic processes is arguably one of the greatest concerns to the scientific community today (Goudie, 2000). The decoupling and individual study of the relative contributions of the two can never be simple. In this study, focused on the flux of sediment from a small catchment, the interrelationship between climate and human activity is essentially one-sided, making the process of establish causative links with catchment denudation much simpler.

It was changing climatic conditions, in the form of flood periodicity, that in part shifted the Reynolds' agricultural focus at Tocal towards pastoralism in the late 19th century (section 4.6.1), which in turn led to an acceleration of soil loss (Figure 8.10). During periods of drought, the disturbance of soils increased due to decreasing vegetation cover and declining organic content. With the return of wetter conditions, the potential erodibility of these soils may have initially been high, but with increasing vegetation cover stabilisation would have occurred. That the greatest level of soil loss from the Tocal catchment (in the 1920s) was contemporaneous with the second highest annual rainfall recorded since the beginning of instrumental records in this region suggests a clear causative association (Figure 8.12). And yet, if the cattle and sheep of the Reynolds family had not devegetated and degraded the catchment slopes, greatly increasing the availability of readily erodible material, would the climate–soil loss connection have existed?

It is likely that years of equally wet conditions prevailed before the arrival of Europeans in the lower Hunter valley. The longest records of floods in Australia are from the nearby Hawkesbury River in western Sydney (Figure 2.1). The flood of 1799 on the Hawkesbury was the second largest on record, and was not surpassed until 1867 when catastrophic flooding inundated the Hawkesbury-Nepean River valley (Figure 8.13). The whole township of Windsor on the banks of the river was destroyed by the 1799 flood, and with it the colony's main source of crops (Nichols, 2001). Interestingly, this flood took place before any human-modification of the upstream Hawkesbury-Nepean catchment area had occurred (the settlement of Sydney had only taken place in the decade before), suggesting that large magnitude floods were a natural feature of the system before

European settlement. The Hawkesbury flood record shows that a period of much wetter conditions prevailed in central eastern Australia in the two decades before European settlement in the Paterson region. Despite the flood-dominated regime, soil loss before the 19th century at Tocal did not exceed the assumed rate of soil formation, and was orders of magnitude lower than the denudation experienced in the early 20th century (Figure 8.12). This would suggest that the influence of climatic shifts on erosion in the undisturbed catchment of Tocal was relative minor.

While the climatic conditions operating in eastern Australia during the late Holocene are not well documented (with either great resolution or certainty [section 3.3]), the detailed observational record since the late 19th century from the Tocal area provides a good indicator of climatic trends, although for a limited period of time. Still, a comparison of this record with the high-resolution sediment yield record from Tocal presents an opportunity to examine what role climatic fluctuations may have had in controlling catchment erosion. Importantly, connections made between climatic events, such as floods or droughts, and the catchment's response during this 150 year period may prove useful in explaining elements of the environmental record for the millennia before European settlement.

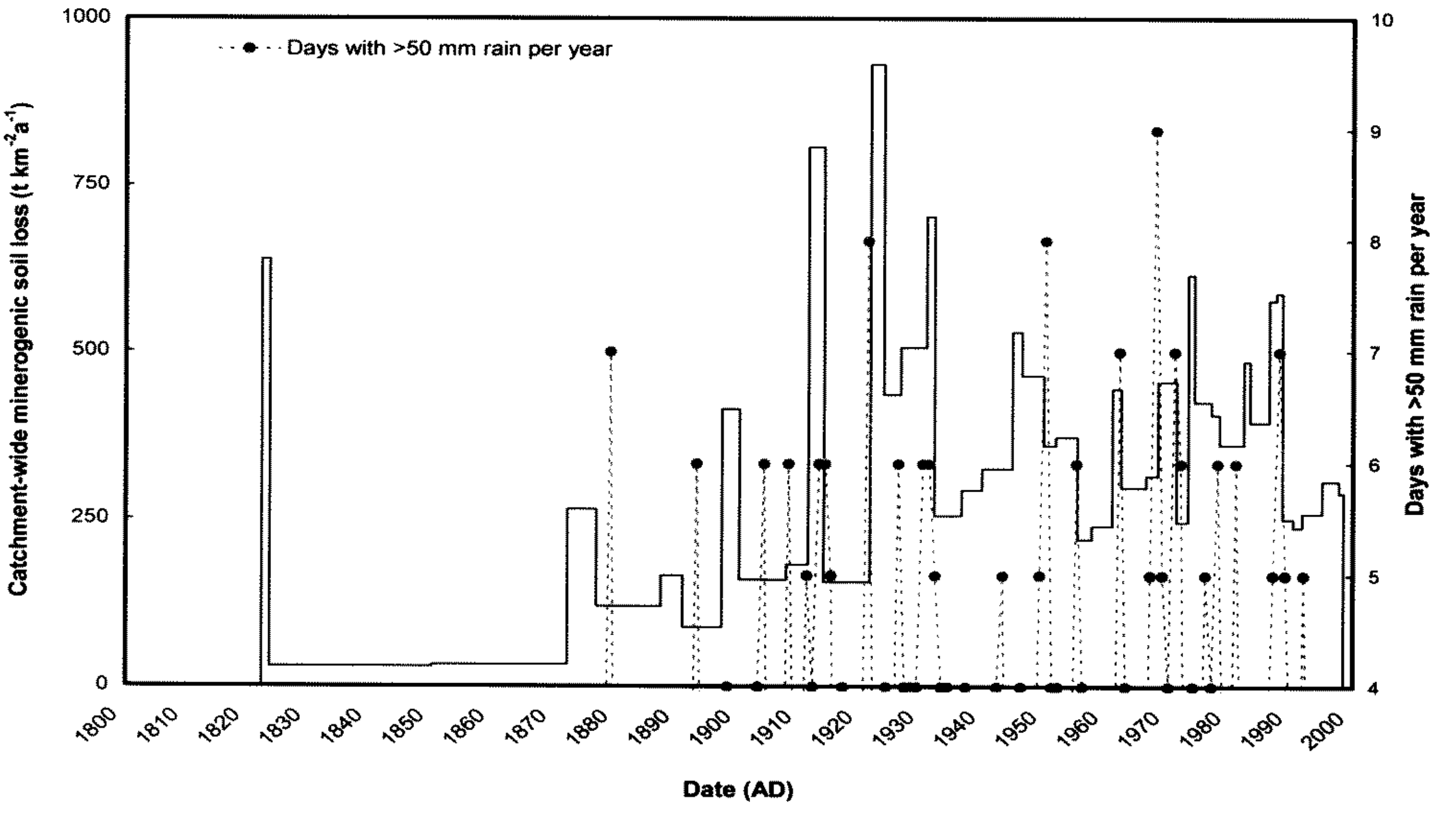


Figure 8.12 A comparison of catchment-wide minerogenic soil loss rates from Tocal Homestead Lagoon with the 140 year proxy flood record of 50 mm initial storm loss. Rainfall data are from the combined West Maitland, Tocal and Tocal Post Office (Paterson) data sets (section 2.6). Unfortunately, no rainfall data are available for the period 1861–1870.

A comparison of the historical soil loss record from Tocal with the proxy flood record for the area suggests that at least on some occasions episodes of high runoff and periods of high sediment flux coincide (Figure 8.12). This coupling seems direct (and short lived) in the early 1910s, the early 1920s, the early 1930s and the late 1960s. Peaks in high runoff years and soil loss are decoupled before 1910 AD, though there are no rainfall data available to examine the relationship from 1861 to 1870.

More evidence of the influence of climatic conditions on catchment sediment yield is provided by the 140 year record of drought conditions at Tocal (Figure 8.14). During some phases of European settlement, prolonged drought conditions have coincided with periods of reduced catchment sediment yield. The evidence of this is most convincing during the 1880s, the 1900s, the mid-1930s to mid-1940s and the early 1990s.

Before *c.* 1880, floods and droughts appear to have had little impact on soil loss. Although only a few years of rainfall data are available for the period before 1870 AD, evidence from elsewhere in the Hunter valley shows that periods of wetter than average and drier than average years occurred throughout the 19th century (Erskine and Warner, 1988).

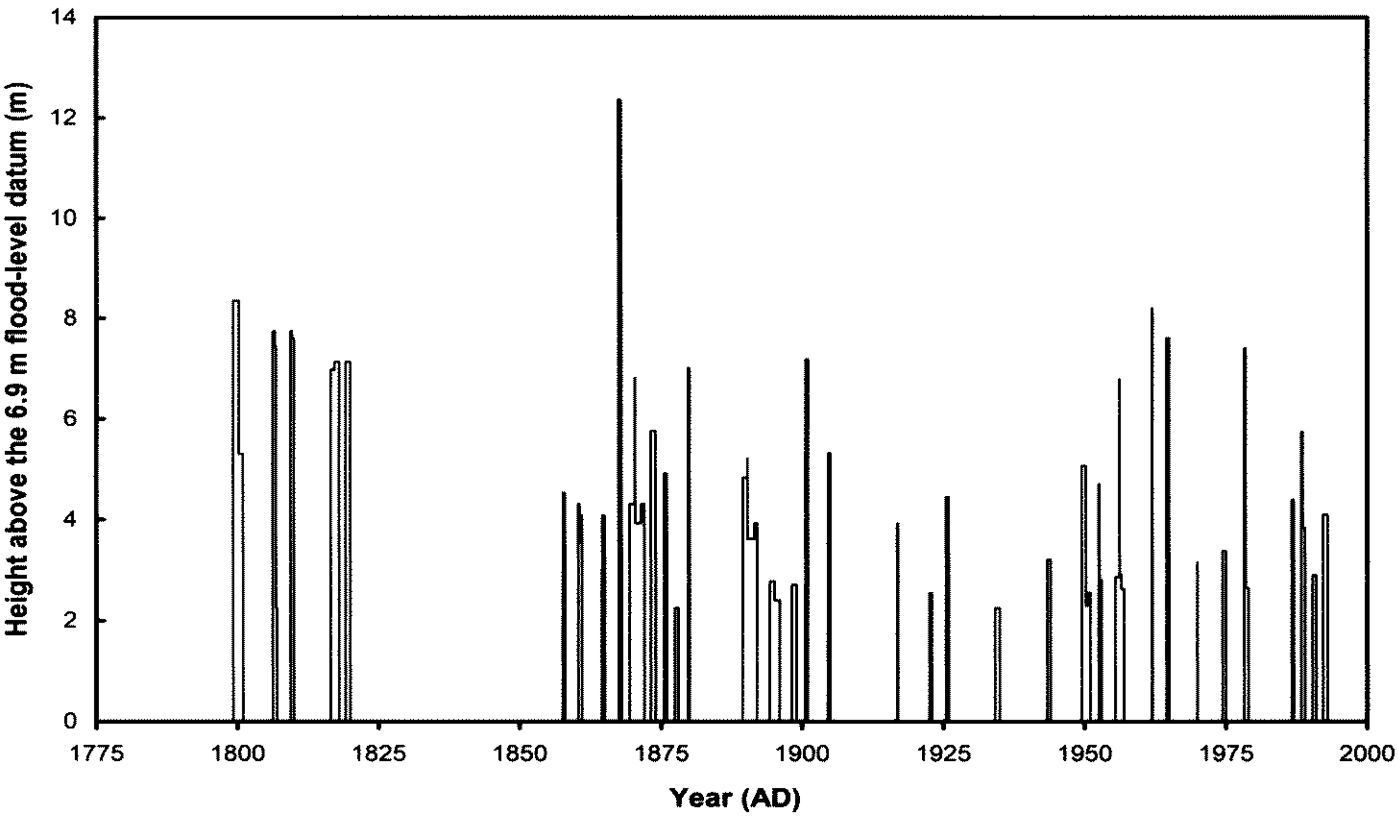


Figure 8.13 Flood events on the Hawkesbury River measured at Windsor Bridge, New South Wales, eastern Australia for the period 1799–1992. Flooding at Windsor Bridge occurs when river stage exceeds 6.9 m. Data from Nichols (2001).

The long record of floods from the Hawkesbury River shows the onset of a flood-dominated regime in 1857 which continued until 1880, during which time the greatest flood since European settlement occurred (1867) (Figure 8.13). Despite these major hydrological perturbations, soil loss remained both low and remarkably stable from *c.* 1830 to *c.* 1880 at Tocal (Figure 8.10), mimicking the long-term soil loss record from the sediment-supply limited catchment of Little Llangothlin Lagoon (Gale and Hoare, 2005). Were similar catchment processes at work at Tocal? An examination of the role of land use in controlling soil loss at Tocal may shed some light on this question.

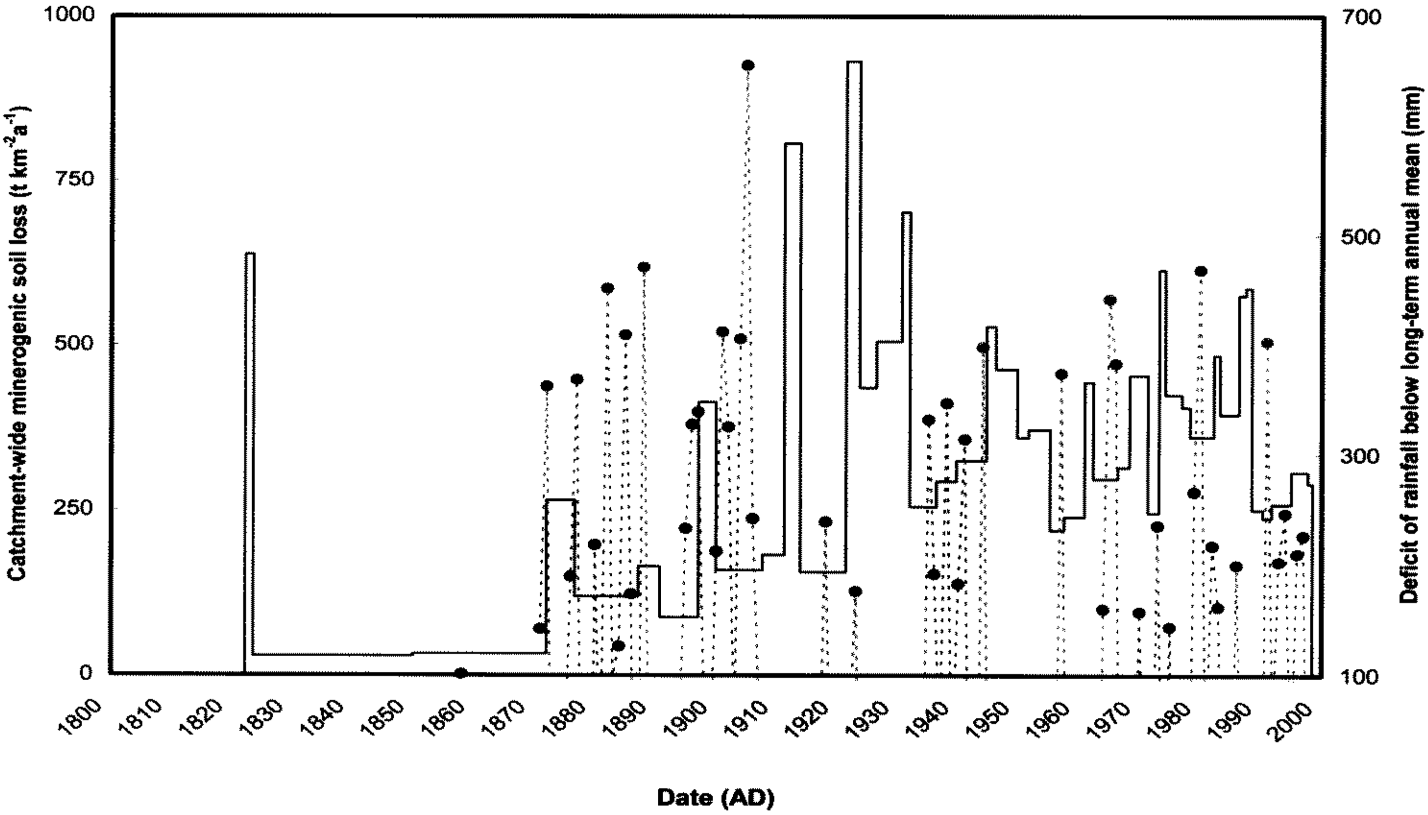


Figure 8.14 A comparison of catchment-wide minerogenic soil loss rates from Tocal Homestead Lagoon with the 140 year proxy record of drought conditions, based on the total annual rainfall (in mm) below the long-term mean. Rainfall data are from the combined West Maitland, Tocal and Tocal Post Office (Paterson) data sets (section 2.6). Unfortunately, no rainfall data are available for the period 1861–1870.

8.3.7 European land use and soil erosion at Tocal

A detailed, long-term examination of the impact of land use on catchment-wide soil loss at Tocal is hampered by two problems. The first is that of gaps in the chronology of agriculture at Tocal, while the second relates to the quality of historical data for some of the years of the post-settlement period. Those gaps in qualitative agricultural data can be seen in Figure 8.15. A more intractable problem, however, is how these changing stock numbers should be interpreted. The data represent total stock numbers for the entire Tocal site, not just for the Tocal Homestead Lagoon catchment. This may be particularly problematic for the period since the 1960s, when the college increased its land holdings, and entire herds of animals were kept outside the catchment during various periods.

The conspicuous 50 year period of low soil loss rates that follows the initial decade of European settlement at Tocal does not coincide with any reduction or cessation of agricultural activity. Indeed, up to 1829, in conjunction with his continued activity in cropping operations, Webber was stocking sheep at extremely high densities at Tocal (Figure 8.15) However, after 1830, Webber was granted an extra 3000 acres of land, probably most of which lay beyond the Tocal Homestead Lagoon catchment (section 4.4.5). The little information that has survived on the location of Webber's agricultural activities suggests that his crops may have been located to the immediate north and northwest of the lake, on the south bank of Webber's Creek (Figure 2.1). However, after 1830, the amount of agricultural activity that took place around the lake cannot be known and much of Webber's stock may have been kept beyond the lake's catchment. This may in part account for the rapid decline in catchment sediment yield shortly after European settlement.

An alternative hypothesis to that of land use control of erosion concerns the availability of readily erodible soil material following the extremely high rates of soil loss that immediately followed settlement (which may have been as high as 1200 t $km^{-1}a^{-1}$ at some locations [section 8.3.5]). Such a prolonged period of consistently low soil loss is typical of a sediment-supply limited system (Gale and Haworth, 2005). More detailed evidence, utilising the careful examination of sediment physical and chemical properties, will be used to examine this thesis in section 8.5.

Following the nearly 50 year period of relatively low soil loss, an increase in catchment erosion by eight times is recorded in *c.* 1870 (Figure 8.16). This increase coincides with one of the most major shifts in land use at Tocal since European settlement. Since the 1840s, Charles Reynolds had been begun to re-orientate his interests from crops to cattle and horses. In the early 1870s, his stock numbers increased rapidly, leading Reynolds to clear the remaining catchment vegetation, and forcing him to acquire additional acres surrounding Tocal to accommodate the swelling numbers. By the turn of the century, catchment-averaged rates of soil loss from the catchment slopes had increased to more than 400 t $km^{-2}a^{-1}$; it is from this period that we have the first photographic evidence of cattle herds grazing on the immediate slopes of Tocal Homestead Lagoon (Figure 4.13).

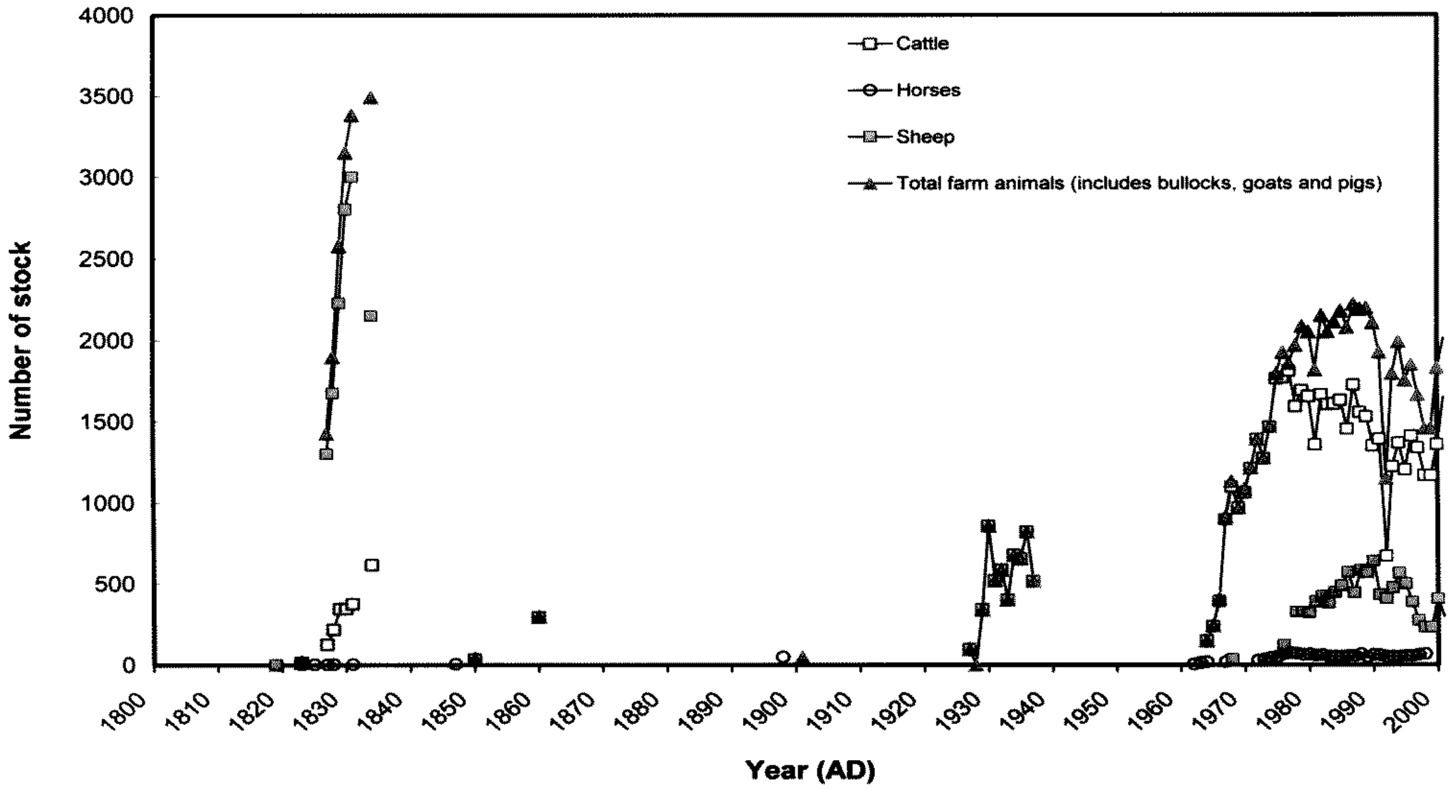
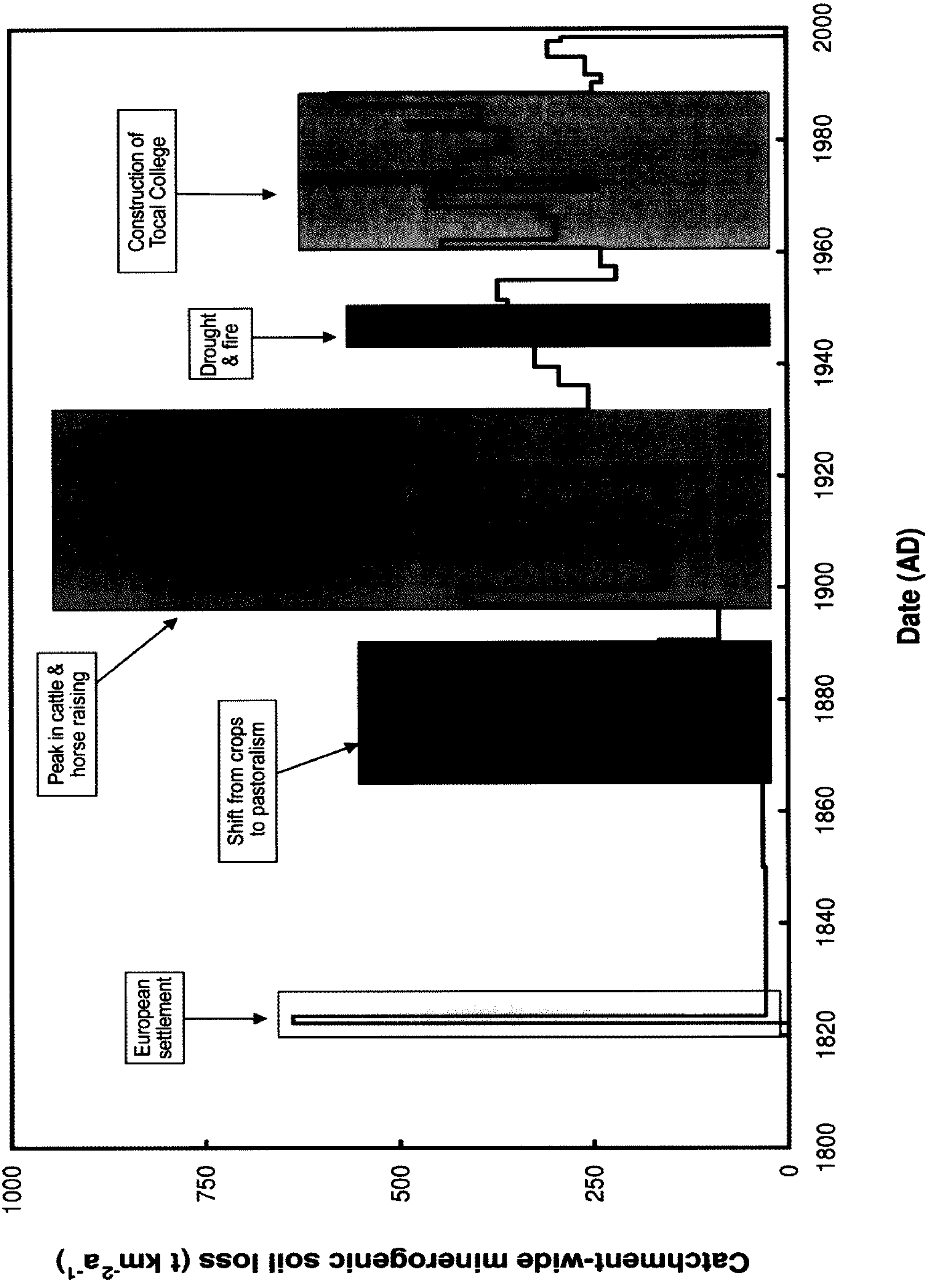


Figure 8.15 Changes in animal populations at Tocal, Paterson, eastern Australia, 1822–1998. Data taken from a wide range of sources cited throughout chapter 4. Reproduced from section 4.10.

Land use in the early decades of the 20th century at Tocal was dominated by cattle and horse rearing. This period coincides with the highest rates of soil loss recorded at Tocal (Figure 8.16). Unlike previous periods, the historical records shows that stock were kept in the catchment of the lake, and some evidence of their impact on vegetation cover and erosion has been presented (section 4.6.3). Although exact stock numbers for this period are unfortunately unavailable, the historical record implies that Tocal was one of the leading cattle enterprises in the country at this time, and that agricultural activity was intensive (section 4.6). There is strong evidence, therefore, to establish a link between cattle ranching at Tocal and the peaks in catchment-wide soil loss for the early 20th century. However, soil loss fluctuates significantly within this *c.* 30 year period; the majority of deposition in the lake took place in brief windows of time, which were separated by longer periods of lower erosion (Figure 8.16). These fluctuations may relate to particularly high-magnitude events in the site's history. The bushfire of 1905 (section 4.6.3) appears to have had a relatively minor impact on soil loss, with rates increasing from 160 to 182 t km^{-2}a^{-1} shortly after the event. This rise is soon afterwards overshadowed by the increase in erosion to 807 t km^{-2}a^{-1} in *c.* 1912, which may record the massive catchment disturbance created by the construction of the North Coast Railway in 1911 (section 4.6.3).


Construction of Tocal College
Drought & fire
Peak in cattle & horse raising
Shift from crops to pastoralism
European settlement
Date (AD)
1800
1820
1840
1860
1880
1900
1920
1940
1960
1980
2000
Catchment-wide minerogenic soil loss (t km^{-2}a^{-1})
0
250
500
750
1000

Finally, the possibility of climatic processes being responsible for the sediment yield fluctuations of the early 1900s cannot be ruled out. The driest calendar year on record at Tocal was 1905 which, despite continued pressure on the soils from intensive agriculture, coincides with a decade of lower erosion rates (Figure 8.16). This suggests that at least at some stages of the long-term erosion record, interactions between climate and land use may have been important (section 8.48, below).

Fire in the Tocal catchment in the mid-1940s coincides with a large increase in catchment erosion, following a period of generally lower erosion associated with declining agricultural activity (section 4.7.1). This contradicts the earlier evidence (based on the fire of 1905) that fires may have had comparatively little impact on soil loss. The 1944 fire is clearly recorded in the sediments above 0.64 m depth in core TCA9b (Figure 6.4), which have been dated to the 1940s by ^{210}Pb methods (section 7.7.4). The fire occurred during an extended dry period at Tocal which probably did not finish until 1949–1950 (Figure 8.12). As such, the increase in erosion from this time cannot be related to climatic shifts, and probably records the impact of the fire on increasing soil loss.

The construction of the Tocal Agricultural College in the early 1960s marked the end of 15 years of negligible agricultural activity. Erosion rates, however, did not reach their minimum until after *c.* 1955, possibly 10 years after farming at Tocal was practically abandoned. It is possible that this may represent some continued instability in the catchment in the years after the cessation of agricultural activity, maybe reflecting the time required for vegetation to re-colonise and stabilise topsoils, or the time for sediment slugs to move through the catchment to the lake. Realistically, however, the observed delay in declining erosion rates probably lies within the range of dates used in these calculations (Table 7.7), and there may be no real 'lag' in erosional response. A clear increase in soil loss occurred in the early 1960s and coincided with the construction of the college, beginning a new phase of elevated erosion rates that continued until the 1990s. The correlation between the two records at this point is not surprising, given both the initial shifts in land use in 1960 (large-scale clearing of the overgrown lake catchment, earthworks and construction) and, over the longer term, the reinstatement of crops, sheep and cattle at the site.

The last decade at Tocal has seen a marked drop in rates of soil loss to values consistently less than 300 t $km^{-2}a^{-1}$. Although relatively drier conditions prevailed through the 1990s, this reduction in soil loss probably reflects those improvements made at Tocal in adopting sustainable agricultural practices since the 1980s (Laffan, 2003). These include the replanting of riparian vegetation, the maintenance of soil cover on slopes, and the identification and remediation of degraded soils.

In summary, there is good evidence that the lag time between modification of the Tocal catchment and corresponding increases in sediment yield has been negligible. Beyond the geomorphic potential of the site to provide direct linkages between the catchment slops and the lake basin, the strongest evidence of rapid sediment delivery comes from the catchment sediment yield record from the early 1820s, when riparian deforestation, catchment clearance and the onset of European agriculture resulted in a massive spike in catchment soil loss that saw erosion rates immediately escalate by more than 210 times. Flood records from the Hawkesbury River show that arrival of Europeans at Tocal in the early 1820s corresponded with the onset of a *c.* 35 year period of drier conditions in central eastern Australia (Figure 8.13), effectively ruling out a climatic cause for the peak in erosion. As the catchment's response to human activity appears to have been quite rapid, direct comparison between the chronology of agricultural activity and the soil loss record since European settlement are probably valid. This is true for at least the early part of the post-settlement record, while further evidence of a direct link between catchment erosion and lake sedimentation comes from the fire and charcoal record in the mid-20th century (section 8.6.4).

During many periods in the 174 year record, shifts in land use at Tocal have coincided with marked changes in catchment sediment flux, and a causative link between the two seems likely for much of the post-settlement period. The influence of catchment land use on soil loss was probably most evident at the beginning of the 20th century, when cattle-based agriculture was at its peak (Figure 8.16). The periods of most severe soil loss were linked to high-density stocking of hard-hoofed animals (mainly sheep and cattle). The impact of cropping activities appears to have been, in comparison, relatively benign. However, the evidence of this is much less clear, being obscured by uncertainties regarding the location of agricultural activity during the 19th century and the possibility of sediment-supply limitation following the severe erosion of the first decades of settlement at Tocal. The importance of bushfires in controlling rates of erosion at Tocal is ambiguous at present. The impact of the 1905 bushfire on the catchment was relatively small, and its direct impact on soil loss was possibly obscured by contemporaneous shifts in climate and rare high-magnitude events such as the construction of the railway line through the catchment. In contrast, the 1944 fire occurred during a period of greatly diminished agricultural activity and stable climatic conditions, suggesting that fire alone may cause significant disruption to soils and may have enhanced erosion rates at Tocal.

8.3.8 The combined impact of human activity and climate on soil loss

Comparing the record of catchment-wide soil loss to Tocal's history of land use has shown that although high-impact agriculture may have persisted over decadal scales (such as from the 1890s to the 1920s), the resultant temporal trend in soil loss appears, at least in part, to have been episodic (Figure 8.16). In the case of the period of Tocal's peak cattle activities at the beginning of the 20th century, soil loss from the catchment was confined to several short periods of extremely high erosion. There is some evidence that the intermittent periods of lower erosion broadly coincided with periods of drier conditions. This is most apparent in the period 1901–1906 (Figure 8.14). Periods of much wetter conditions largely correspond with the spikes in soil erosion in the early 1910s and early 1920s (Figure 8.12). The effect of decreased rainfall in combination with sustained agricultural activity would have decreased an already depleted vegetation cover, leading to even greater disruption of soil profiles. However, under the severe drought conditions of the early 1900s, the transport component of the erosion cycle may have been temporarily impeded, leading to relatively low rates of deposition in the lake.

If the thesis of transport-controlled sediment-supply is pursued, then the reinstatement of wetter conditions while agricultural activity was maintained would have resulted in the rapid delivery of eroded material into the lake. This material may have been stored temporarily on catchment slopes during rare high-magnitude droughts (such as those of 1904–1906), and transported rapidly from poorly vegetated slopes into the lake. Although the idea of transport limitation contradicts to some extent the model of direct linkage from the catchment to the lake, if such limitation occurred since European settlement then it was confined to discrete periods of hydrological extremes. Under such conditions, even the most efficient sediment delivery network may be interrupted. For much of the 174 year post-settlement record, there is good evidence that rapid and direct sediment transport to the lake took place.

Beyond the first three decades of the 20th century, the only other period when transport limitation may have operated is 1960–1970 (the first decade of the Tocal Agricultural College) (Figure 8.12). In this period of increased and sustained farming, fluctuations in soil erosion are broadly synchronous with flood and drought conditions. However, these short-term variations in erosion could equally be a consequence of uncertainties in the timing and location of agricultural activity after the college's construction. As the college greatly increased its size after the mid-1960s, it is feasible that at some stages hard-hoofed animals or crops may have been relocated beyond the lake's catchment. This is even more likely if signs of erosion (as suggested by Figure 8.7) were acted on.

8.4 Environmental history of Tocal Homestead Lagoon

8.4.1 Introduction

Having discussed the record of lake sedimentation, the parallel record of sediment chemical, magnetic and sedimentological properties can be considered with the aim of providing a detailed account of environmental change during the last 2000 years at the site.

The most distinct change in the 2.00 m lake sediment sequence is that which accompanied the arrival of Europeans in the catchment in the 1820s. Unit 1 is thus defined as that spanning the environmental record for the *c.* 1900 years before European settlement (2.00–1.26 m), while Unit 2 encompasses those sediments deposited since then (1.26–0.00 m). These units have been further subdivided based upon major shifts in magnetic, chemical or sedimentary properties (or sometimes all three). In the case of Unit 1, not all of these sediments have been individually dated, and the chronology is largely based upon linear interpolations between several ^{14}C dates. Within Unit 2, by contrast, the majority of the sequence has been directly dated, so that ages are known with confidence.

8.4.2 Unit 1: Pre-European erosion and sediment provenance at Tocal

The relative absence of fine secondary ferrimagnetic grains in Unit 1 (zero χ_{fd} and 'hard' S-ratios), particularly when compared with the uppermost sediments in the core, would suggest that these sediments were not derived from topsoils. Alternatively, the sediments may formerly have contained soil-derived ferrimagnetic material, which has undergone post-depositional modification (e.g., dissolution) of magnetic minerals (section 6.11.1.4). That the pre-European sediments are gleyed and generally mottled throughout lends some support to this argument. Gleyed sediments are characterised by zero or very low values of χ_{fd} as the finest magnetic grains dissolve during the gleying process (Gale and Hoare, 1991, p. 214). The dissolution of soil-derived magnetite was reported by Erikson and Sandgren (1999) in the pre-contact sedimentary record from Lake Haubi, Tanzania. This was explained as being in part due to very low sedimentation rates and gleying (periodic wetting and drying) of the sediments.

Although the study of the provenance of pre-European sediments at Tocal was not definitive in its conclusions (section 6.8.5), there is strong evidence that these lake sediments were derived from soils similar to the pre-European soil preserved beneath Webber's cottage. This soil (TCWC1) is chemically and physically distinct from modern catchment soils, being poorly structured, friable and highly erodible (section 6.9). The accelerated erosion of soils such as this from the catchment through the 19th and 20th centuries may in part explain the absence of any identical soil types at Tocal today. The high-erodibility of pre-European topsoils meant that they probably were quickly removed from the catchment, leaving only the more resilient, deeper soils. In addition to erosion, the traffic of hard-hoofed beasts, and the use of ploughs and heavier machinery would have had led to the compaction of the original soil profiles. Today, truncated and compacted soils are almost all that exist at Tocal (section 2.3.3).

8.4.3 Unit 1.1: 2.00 m to 1.54 m (*c.* 200 BC to *c.* 800 AD)

What was Tocal Homestead Lagoon like *c.* 2000 years ago? It is clear from the environmental magnetic record in the lower sediments at Tocal that a very different soil erosion and sedimentation regime operated in the catchment throughout the *c.* 1900 years before European settlement (section 6.11.1). The rate of deposition of the magnetically weaker fine-grained silts in the lake was low (*c.* 0.3 kg $m^{-2}a^{-1}$), which suggests that this sub-unit was deposited during a period of relatively stable conditions. While most properties remain stable in the lowermost sediments, elevated LOI_{430} values were recorded between 1.98 m and 1.74 m (*c.* 200 BC to *c.* 400 AD), though transitions into and out of this sub-unit make its exact boundaries difficult to define (Figure 6.5). The sediments from this period are heavily mottled, which has probably discouraged the preservation of macrobotanical remains. This sub-unit, characterised by higher plant organic content and mottling of the sediments is indicative of lower, fluctuating lake levels and higher organic productivity in a shallow lake. This suggests that the period *c.* 200 BC to *c.* 400 AD may have been characterised by relatively dry conditions.

What information do other palaeoenvironmental studies from the region provide on the environmental conditions *c.* 2000 years ago? Several sites from eastern Australia suggest that the pre-contact period was characterised by relatively stable environmental conditions (Gale *et al.*, 2004). The closest palaeoenvironmental records, from Barrington Tops (Figure 2.2), suggest that vegetation shifts during the last 2000 years were minor compared to those of the entire Holocene. However, these sites are thought to have been relatively insensitive to minor climatic perturbations during the late Holocene (Dodson *et al.*, 1986). In the southern Hunter valley, pre-European riparian vegetation on upper Wollombi Creek is thought to have been characterised by wet sclerophyll forest, with some rainforest elements and *Casuarina* represented in the community (Bennett and Mooney, 2003). A similar vegetation community may have existed at Tocal during this period. However, a notable absence from the upper Wollombi Creek vegetation reconstruction is red cedar (*Toona ciliata*), which historical accounts show dominated river banks in the Hunter valley (section 4.4). The disparity between the historical and palynological record of red cedar at Tocal suggests that its absence from pollen records is most probably a product of its pollination mechanism, rather than problems with the historical record. Preliminary results from the Tocal pollen record show little significant variation during the *c.* 1900 years before European settlement (Dr D.A. Penny, personal communication, 2005), lending support to the argument that vegetation was relatively insensitive to smaller environmental shifts during the late Holocene.

Macroscopic charcoal is near ubiquitous throughout the Tocal sedimentary record before the point of European contact. Two brief exceptions are at the very base of the sequence, 2.00 to 1.98 m, and between 1.74 and 1.72 m. Constant, low levels of charcoal in Australian sediments have been interpreted as recording continual Aboriginal occupation (Clark, 1983; Head, 1988), although this does not take into account the natural background incidence of fire in the Australian environment. Macroscopic charcoal in lake sediments is generally thought to have its source in local fires that burned within the drainage basin (Clark, 1983; MacDonald *et al.*, 1991). The macroscopic charcoal in Unit 1, therefore, probably records catchment-scale fires at Tocal. Whether the frequency and/or intensity of firing in the Tocal landscape had a substantial impact on catchment vegetation and erosion cannot be easily tested; if fire at Tocal were as regular as the charcoal record suggests, then some state of equilibrium may have been attained.

Whether this record of continual burning represents human activity is also difficult to test. Firing of the Australian landscape by Aboriginal people was observed by explorers centuries before English settlement began in the late 18th century (Crowley and Garnett, 2001). At Tocal there is strong evidence of Aboriginal land use in the lake's catchment before European settlement (section 4.2.4), while the earliest historical observations of Aboriginal people in the Hunter region and elsewhere in eastern Australia (from 1770) document their burning of the landscape (section 4.2). Considering these separate lines of evidence, the pre-European record of burning at Tocal potentially records Aboriginal human activity as early as *c.* 200 BC.

Most archaeological sites in the Hunter valley date from the last 2000 years (section 4.2). The macroscopic charcoal record for this period, therefore, could be interpreted as a record of Aboriginal land use either at Tocal, or in the surrounding area, dating from *c.* 200 BC. The steady increases in Mg, K, Ca and Mn in Unit 1.1 suggest that catchment disturbance slowly, but consistently increased through this period as might be expected following increasing human activity (Mackereth, 1966; Engstom and Wright, 1984; Foster and Lees, 1999) (see also section 6.11.2). A very similar trend is recorded in the SIRM/χ_{lf} quotient (Figure 6.13), showing that increasing amounts of fine-grained magnetic minerals, typical of surface soils (Thompson and Oldfield, 1986), were eroded into the lake during this period. These separate lines of evidence are suggestive of a steady, thought relatively minor, increase in catchment disturbance, as would accompany increasing human activity at Tocal. However, without more data, no conclusive arguments can be made for this case at present.

In summary, from *c.* 200 BC to *c.* 800 AD, lake sedimentation at Tocal was low, with distal locations in the lake receiving little or no infill. Frequent fires took place in the catchment. However, the lake sediment record indicates only minor increases in catchment disturbance, which may be explained by other processes, such as human activity. Catchment conditions from *c.* 200 BC to *c.* 800 AD were arguably very stable; the catchment-wide erosion rate during this period was 23 t $km^{-2}a^{-1}$. Modern observations at Tocal show that rates of catchment-wide soil loss of the order of 250 t $km^{-2}a^{-1}$ (Figure 8.10) may be tolerated without catchment instability or gullying. Vegetated soils deeper than those found today characterised the surrounding catchment (section 2.3.3), probably without any gullying or significant drainage incision. Such soils were weakly aggregated and uncompacted, possessed relatively high infiltration capacity, and, as demonstrated in section 6.9, were highly erodible (see Gale [2003] for more evidence of this). Evidence from the Tocal lake record has suggested that climate in the region was relatively dry, with low and fluctuating lake levels after *c.* 200 BC encouraging organic productivity. A return to lower organic productivity in the lake may date from *c.* 800 AD, but certainly from some time after *c.* 50 BC (the upper age limit on the ^{14}C date from 1.94–1.92 m).

8.4.4 Unit 1.2: 1.54 m to 1.34 m (*c.* 800 AD to *c.* 1200 AD)

The base of this sub-unit marks a stratigraphic boundary in the Tocal lake sequence, with those sediments above this depth being characterised by prominent laminae (Figure 6.3). Sometime around 800 AD, the amount of P in the sediments of Tocal Homestead Lagoon began to rise steadily above relatively stable background conditions (Figure 6.21). The sediments in Unit 1.2 display no sign of mixing or bioturbation, being well laminated and possessing very few plant remains (Figure 6.3). As other possible explanations for such an increase in P have been tested and discounted, (section 6.11.2.2), the P increase is considered to be contemporary with the sediments.

The only two environmental pathways for increased P in the lake at this time are enhanced detrital P input into the system, indicative of increased catchment weathering or erosion, or an increase in human and/or animal activity in the catchment (section 6.11.2.2). Due to the coarse chronological resolution of the lower portion of the sequence, it is impossible at present to detect with any certainty fluctuations in catchment sediment yield that may have occurred through this period. What is known, however, is that the mean rate of soil loss until *c.* 1290 AD was low (23 t $km^{-1}a^{-1}$) (Figure 8.11) and, as discussed earlier (section 8.3.5), erosion was most probably less than soil formation rates on the alluvial parent material at this time. It is therefore unlikely that any significant increase in the intensity of catchment erosion, increasing the flux of P from soils, marked this period. Also, comparisons of the increase in P above 1.54 m with the behaviour of Ca, K, Mg and Na in Unit 1 suggest that the P rise was independent of either increased erosion or soil weathering (section 6.11.2.2). Considering the inadequacy of any other mechanism to explain the P rise above background at 1.54 m, the only remaining explanation for the rise in P is related to human activity. As such, it is cautiously suggested that from *c.* 800 AD, human activity either began or significantly increased in the catchment.

There is no historical information to show whether or not Aboriginal people were living at Tocal when Webber arrived in *c.* 1820. There is overwhelming evidence, however, that significant numbers of Aborigines had in the past lived within the catchment of the lake (section 4.2.4). There are also accounts of areas around Tocal being inhabited by Aboriginal people at the time of European settlement in the early 1820s (section 4.2.3). It is thus not inconceivable that the date *c.* 800 AD marks the initiation of Aboriginal occupation at Tocal, though such an assertion can be no more than speculative at present.

8.4.5 The beginning of Aboriginal occupation in the lower Hunter valley

How does the evidence of increasing human activity at Tocal from *c.* 800 AD fit with what is currently known about the Aboriginal history of the region? Although the oldest archaeological site in the Hunter valley dates from *c.* 11 000 cal. BP, most sites suggest that intensification of human activity did not occur until the late Holocene, with the majority of sites dating from *c.* 2000 cal. BP (section 4.2.2). From this perspective alone, the hypothesis of Aboriginal occupation at Tocal not beginning until after *c.* 1200 cal. BP seems unlikely—especially considering the attractions of the site for occupation—but not improbable.

The deposition of charcoal in Boggy Swamp on Barrington Tops (50 km northeast of Tocal, Figure 2.2) appears to have been regular since *c.* 12 000 cal. BP (Dodson *et al.*, 1986), albeit at low levels. Charcoal densities in sediments in the Barrington Tops area increased drastically after 3000 cal. BP (*c.* 2000 cal. BP in Boggy Swamp), but whether this is due to increased Aboriginal burning or climatic processes is unclear. There are no records of lake sediment chemistry from the region that can be compared with that from Tocal. The Tocal record therefore stands alone in providing speculative evidence of increasing human activity during this period. What do other environmental records from Tocal suggest about possible human activity from 800 AD onwards? The charcoal record suggests the amount of burning in the area of the lake during this period was probably little different from that of the previous centuries. Quantitative analysis of the charcoal record at Tocal however, will be necessary to provide more conclusive information about the history of fire during this period.

8.4.6 A climatic optimum at Tocal recorded from *c.* 900 AD?

A pronounced decline in the S-ratio beginning at *c.* 1.50 m in the sediments of core TCA9b (*c.* 900 AD) cannot be explained other than by a major decline in ferrimagnetic mineral content in the Tocal sediments, probably due to decreased catchment erosion (section 6.11.1.3 and Figure 6.13). The change in the S-ratio suggests that the shift was not rapid, but probably a result of a gradual transition to a period of reduced catchment erosion. The decrease in ferrimagnetic mineral concentration may also be due in part to magnetic dissolution in these gleyed sediments, with zero χ_{fd} values until 1.36 m depth providing support for this thesis. However, the pattern of zero χ_{fd} begins at 1.70 m, suggesting that if magnetic dissolution were responsible, then such processes had been important before the dramatic shift in the S-ratio began. It is tentatively proposed then that a significant reduction in the erosion of soil-derived magnetic minerals to the lake is recorded in the sediments above *c.* 1.50 m at Tocal. No sediments in Unit 1.2 (1.54 m to 1.34 m) have been directly dated, and only an estimate of the period recorded by these sediments can be provided: *c.* 900 AD to *c.* 1200 AD. The upper age of this feature can be constrained, however, with confidence. The trough in the S-ratio is centred on sediments at 1.36–1.34 m, with increases in the S-ratio taking place by 1230 AD (the lower age limit of the ^{14}C date from 1.32–1.30 m).

The appearance of much finer grained, light grey silt during this period (Figure 6.3) suggests the possible onset of more quiescent erosional conditions at Tocal, and by extension, provides some support for the argument of increased catchment stability after *c.* 900 AD. The plant organic content of the sediments varies only marginally through this period, and much less so than during the period of *c.* 200 BC to *c.* 400 AD (Figure 6.5). Further support for decreased erosion during this period comes from the calculated catchment-wide rate of soil loss in the overlying sub-unit (1.32–1.26 m), which is directly dated to the period *c.* 1250 to *c.* 1500 AD (section 8.5.7, Figure 8.10). The sediments immediately overlying the S-ratio trough record a drop in catchment erosion by more than half, to 10 t $km^{-2}a^{-1}$ (Figure 8.11). Considering the coarse resolution of the erosion rate determined from 1.94–1.32 m, it is conceivable that such a long-term average may mask a much lower rate in the sediments from *c.* 1.52–1.34 m, perhaps similar to that recorded in the adjacent overlying unit, which is also characterised by low S-ratios.

Two theories may be advanced to explain the prolonged period of very low sedimentation at Tocal at this time. First, a possible decline in human activity at Tocal, perhaps part of some broader pattern of disruption in Aboriginal society, may have occurred after *c.* 900 AD. There is little evidence of this in the Tocal sediments, however. No corresponding decline in human activity markers, such as P content and macroscopic charcoal, is recorded through this period (though these indicators alone cannot conclusively prove whether or not such a decline took place). Indeed the opposite is suggested: concentrations of these indicators increase above *c.* 1.50 m, suggesting a possible paradox of an increase in human activity during a period of reduced catchment erosion.

The second, and more defensible theory, is that increasingly warmer conditions may be recorded in the sediments above *c.* 1.50 m, resulting in increased catchment vegetation and stability and decreased erosion. Warmer conditions at this time at Tocal would broadly be coincident with the Medieval Warm Period (section 2.3.2). Preliminary results from the Tocal pollen record for this period show that fern species disappeared from the site during this period (Dr. Dan Penny, personal communication, 2005), lending further support to the theory of a possible climatic shift to warmer, drier conditions. As there is some evidence of increasing human activity in the catchment during this period, *increased* catchment stability would almost certainly have required an increased stabilisation of the catchment surface as would accompany increasing vegetation, and thus warmer conditions. This suggests that the imprint of prehistoric human activity (arguably minor in any case) during this period was small compared to the role of major climatic shifts.

The evidence for the presence of a Medieval Warm Period in the Australian region is reviewed in section 2.3.2. Cook *et al.* (2000) have shown that a long warm period was recorded from 900 AD to 1500 AD in Tasmania, though fluctuations within this period are evident (Cook *et al.*, 1991). Although there is little evidence of warmer conditions in mainland Australia during this period, the significant decrease in erosion at Tocal dates from this time and, in the absence of any other appropriate explanation, is likely to be connected to climatic change, most probably a shift to warmer conditions. The period of possibly warmer conditions at Tocal from *c.* 900 AD to *c.* 1200 AD corresponds well with evidence from temperate locations in New Zealand of a Medieval Warm Period (section 2.3.2), and to the climatic simulations for the southern hemisphere for this period (Goosse *et al.*, 2003).

8.4.7 Unit 1.2: 1.34 to 1.26 m (*c.* 1200 AD to *c.* 1820 AD)

Radiocarbon and palaeomagnetic dating shows that sediments in this unit were deposited between *c.* 1200 AD and *c.* 1820 AD. The lowermost sediment in the sub-unit is no younger than 1160 AD as the sediments from 1.32–1.30 m have been dated to 1290 ± 60 AD (Table 7.5), while the upper limits of the sub-unit underlie the point of European contact in the early 1820s AD, dated using ^{210}Pb and palaeomagnetic methods (section 7.8). Above 1.34 m, several magnetic quotients begin to increase. The base of this unit lies within the S-ratio minimum at 1.36–1.34 m (Figure 6.13), suggesting that the presumed warmer conditions identified in the underlying sediments may have continued during the deposition of the base of this sub-unit. So when did the phase of warmer conditions suggested to have been operating at Tocal before *c.* 1200 AD finish, and catchment stability begin to deteriorate? Based on the S-ratio, catchment instability capable of contributing pedogenic ferrimagnetic material to the basin—overturning environmental conditions of low sedimentation and possible magnetic dissolution—probably did not return until above 1.28 m. By this time (*c.* 1500 AD), S-ratio values similar to those recorded earlier in the sequence had returned. The incidence of gleyed fine-grained silts observed above 1.54 m (*c.* 800 AD) is much less frequent above *c.* 1.32 m (1290 ± 60 AD). The sediments above this depth are composed of coarser silts, providing further support for increasing catchment disturbance. If the Tocal sediments do record the Medieval Warm Period, then the evidence that warmer conditions had finished by *c.* 1500 AD is in agreement with the date of the close of this period elsewhere in the southern hemisphere (section 3.3.3).

Charcoal content increases in this unit to some of this highest values found in the entire sequence, comparable with those associated with the greatest fires recorded in the post-settlement era (Figure 6.4 and section 4.7.2). Fragments of charred wood of up to 5 mm in length suggest that burning must have occurred in the immediate catchment. The P content of the sediments increases at an even greater rate than in the underlying unit, while chemical indicators of erosion (parallel increases in Ca, K, Mg and Na) also increase (section 6.11.2.1–6.11.2.4 and Figure 6.17). It is possible that human activity, for which there is evidence at Tocal from at least *c.* 800 AD, increased in the *c.* 600 years before European settlement.

While several indicators suggest further increases in human activity in the catchment during this period, and substantial burning of the site, the catchment-wide rates of erosion were the lowest recorded in the 2100 year lake sediment history. From *c.* 1300 to *c.* 1500 AD (1.32–1.28 m), soil loss was 10 t $km^{-2}a^{-1}$, which declined further from *c.* 1500 AD to *c.* 1820 AD to 3 t $km^{-2}a^{-1}$. The error bars on the median ^{14}C age determined for sediments at 1.28–1.26 m depth are quite large, as they take into account the multiple age determinations made on this sample depth (section 7.5). Therefore, the erosion rate determined for the period *c.* 1500 AD to *c.* 1820 AD may be an underestimate of the true rate of soil loss for this period. Palaeomagnetic dating, for example, suggests that the median age of 1480 AD made on two samples from 1.28–1.26 m may be up to two centuries too old. However, as discussed in section 7.5.2, the palaeomagnetic dating is identical to the ^{14}C dates at 95% confidence limits for this depth.

Further increases in catchment disturbance date from *c.* 1500 AD, with major changes in the chemical and magnetic properties of the sediments recorded at this depth (1.28–1.26 m) (Figure 6.18 and 6.19). It is possible that some of these changes may be an artefact of the sampling interval used, with the top of the 1.28–1.26 m sample overlapping the basal sediments deposited during the initial stages of European settlement. However, bearing in mind the evidence of possible Aboriginal occupation at Tocal in the millennia before European settlement, these marked changes in the sediments above 1.28 m depth may also record increasing pre-European human activity.

8.5 Unit 2: 1.26 m to 0.00 m (*c.* 1820 AD to 1998 AD)

8.5.1 Unit 2.1: 1.26 m to 1.18 m (*c.* 1820 AD to *c.* 1870 AD)

The arrival of Europeans at Tocal resulted in a dramatic and near-instantaneous response from the catchment (Figure 8.16). The immediacy of their environmental impact was also recorded elsewhere in the lower Hunter valley. Cunningham (1827), for example, noted increasingly impassable gullies along roadways by 1827 (section 4.4.5). From perhaps as early as 1818, the riparian zone to the north and east of Tocal Homestead Lagoon had been logged for red cedar and, following Webber's arrival in 1822, further catchment vegetation was cleared to make way for crops and stock. The majority of the magnetic proxies increase sharply above 1.26 m, reflecting the sudden influx of soil-derived ferrimagnetic minerals into the basin.

The concentration of P in the sediments continues to increase after European contact at a similar rate to that recorded in the preceding centuries. The P/Fe ratio, however, increases sharply after contact (Figure 6.21), reflecting the greater contribution of non-detrital phosphorus to the lake (section 6.11.2.2). Loss-on-ignition values increase rapidly after European settlement, more than doubling the pre-European level by 1872. In conjunction with the increases in P above 1.26 m, the increase in organic content in the sediments is likely to record the onset of lake eutrophication. An increase in eutrophic conditions in the lake would have promoted the growth of aquatic vegetation in the lake, increasing LOI_{430} values in the unit and increasing sediment porosity (section 6.2.2). Similar changes have been reported following European settlement at other sites in New South Wales (Gale and Haworth, 2002).

The increasing productivity of the lake is also recorded in a massive rise in sulphur in the decades following European settlement. Sulphur levels in the sediments above 1.26 m more than double, mimicking the response of phosphorus to European settlement. This is most likely to be the result of an increase in organic productivity in the lake, suggesting that the lake became more eutrophic (section 6.11.2.5).

Although there is little account of it in the historical record, firing of the environment continued once Webber had arrived at Tocal. In the early years of settlement, fire would have been used to burn tree stumps and remove grasses and shrubs, while bushfires burnt the catchment at least once during this period (in 1835 [section 4.5.1]). Macroscopic charcoal was deposited in the lake continually from contact until *c.* 1872. Although there is no historical record of burning at Tocal in the first *c.* 50 years of European settlement, the charcoal record suggests that regular firing of the site continued until long after settlement. European settlers are known to have used frequent, low-intensity fires to maintain pastures in Australia (Bowman, 2000). An alternative explanation is that following the cessation of fires after the first decades of settlement, charcoal remaining on the catchment's surface may have continued to be eroded in the lake. The end of charcoal deposition in *c.* 1872 may thus represent the final depletion of charcoal from the catchment slopes. Evidence from other palaeoenvironmental records in the region, however, favours the European burning hypothesis. Although charcoal concentrations gradually decline in the sediments of nearby Burraga Swamp in the last two centuries, macroscopic charcoal was consistently deposited in the swamp until the top of the sequence (Dodson *et al.*, 1994).

The onset of European agriculture at Tocal resulted in a massive increase in sediment yield, similar in magnitude to that recorded at other sites in eastern Australia. These high rates were not maintained, however, and despite increasing agricultural activity (Figure 4.24), erosion rates equally rapidly declined to levels only four times greater than those of immediately pre-contact times (Figure 8.10). Another striking feature of this record is the duration of the phase of low erosion, which extended for nearly 50 years. One explanation of this may be that initial disturbance was restricted to a small part of the catchment that was then either stabilised or depleted of soil before larger tracts of the catchment were taken into use. By Webber's own admission, his land could not support the *c.* 2500 animals he kept by 1829. Sheep numbers alone increased to 3000 in the next two years, but with his additional land grants totalling 3000 acres in 1830–1831, stocking densities around the lake probably dropped as Webber's beasts were dispersed to lands north and west of the catchment (section 4.4). Non-detrital P (P/Fe) input into the lake (Figure 6.21), however, continued to increase rapidly through this period, suggesting that no major decrease in animal populations in the catchment took place before the 1870s.

Alternatively, uncompacted, easily disturbed, pre-contact topsoils may have been rapidly stripped from the hillslopes, leaving less erodible lower horizons that limited the supply of material to the basin. The process of supply-limitation of soil loss following European settlement in Australia has been documented by Gale and Haworth (2005). Extensive erosion of duplex soil profiles in the catchment has led to truncated A-horizons, which still persist today, providing support for this thesis. The majority of soil material eroded into the lake following European settlement most likely originated from the slopes south and east of the lake (section 8.2.3.2). Geochemically, these soils were more similar to the pre-European soil from beneath Webber's cottage than to the profiles found today (Table 6.3).

The high erodibility of these pre-European soils (section 6.9) would have greatly contributed to their swift erosion following first contact, limiting the supply of readily erodible material. Perhaps it was the level of denudation of topsoils around the lake by 1829 that led to Webber's statement concerning overstocking on his farm (section 4.4.4)? Finally, it is pertinent to note that climatic fluctuations before 1870 AD appear to have had no impact at all on the low and constant rate of soil erosion from the catchment (Figures 8.12 and 8.14). Although the proxy flood and drought records extend only to the decade before 1870, an inspection of the much longer flood record from the Hawkesbury River shows a return to much wetter conditions in the late 1850s in central eastern Australia (Figure 8.13), with little impact on rates of erosion at Tocal. The apparent lack of influence of climatic conditions in this period lends further support to the sediment supply limitation argument.

8.5.2 Unit 2.2: 1.18 m to 0.94 m (*c.* 1870 AD to *c.* 1912 AD)

It was around 1870 that the Reynolds family began to shift the agricultural focus at Tocal away from crops towards pastoralism (Figure 8.15). The result was increased soil erosion. The sharp contact between units 2.2 and 2.1 and the associated change in lithology marks the influx of sediment after *c.* 1870, possibly coinciding with a return to wetter conditions from the late 1860s (Figure 8.13). Rates of soil loss increased by nearly 10 times to 264 t $km^{-1}a^{-1}$. For the first time in the Tocal record, material eroded into the lake included small amounts of fine sand, suggesting increasing erosion intensity and a possible shift in sediment source from the alluvial soils immediately surrounding the lake to those soils formed on sandstone in the southeastern sector of the catchment. Sand in the duplex kurosols of the catchment is found predominantly at the base of the topsoil layer, immediately overlying the clay-rich B-horizon. The appearance of sands in the lake sediment in this period may have been the result of the erosion of the loamy upper topsoil to a depth that exposed this coarser material. Modern observations of soils in the catchment suggest that at least 0.5 m of soil has been removed since the early 19th century (section 2.3.3), lending support to this hypothesis.

The first photographic record of agriculture at Tocal dates from *c.* 1900, showing Reynolds' cattle and horses grazing the immediate slopes of the lake and the western interfluve in paddocks near Tocal Homestead (Figures 4.12, 4.14–4.16). These images also show that the majority of the southwestern catchment had been cleared of vegetation. *Eucalyptus* pollen in the sequence appears to suffer a decline in abundance after *c.* 1870, which may record the period when the remaining woodland in the catchment and its surrounds was cleared as part of Reynolds' agricultural expansion.

A shift in magnetic mineralogy is recorded in this unit, to sediments characterised by relatively coarse-grained ferrimagnets with high χ_{lf} and zero χ_{fd} values (Dearing *et al.*, 1996). The magnetic assemblage of this unit is typical of less-weathered subsoils (Thompson and Oldfield, 1986; Gale and Hoare, 1991) that may have been exposed to erosion following the stripping of topsoils in previous decades (section 8.6.1). Both the sedimentological and mineral magnetic evidence suggest that erosion of coarser material from the catchment's slopes characterised this period at Tocal. This material may have included subsoil of lower erodibility, suggesting that catchment erosion was relatively intensive. This would be expected considering the increase in pastoral activity during this period (section 4.6.2). Most magnetic measures are remarkably stable during this period (χ_{lf} increases marginally above 1.06 m [*c.* 1897]), taking into account the evidence of enhanced catchment erosion in this period. The increasing plant organic content of the sediments (which reaches its maximum for the entire core in this unit [29% at 1.00–0.98 m, Figure 6.5]) may have had a dilutionary effect on sediment magnetism due to its diamagnetic properties (Dearing, 1999a), and may mask magnetic evidence of the intensification of erosion. Correcting the sediments from 1.00–0.98 m for their LOI_{430} content (using the value of -0.009 x 10^{-6} m^3kg^{-1} provided by Dearing [1999a]) increases their χ_{lf} from 14.45 to 20.78 x 10^{-8} m^3kg^{-1}. This confirms the model of significant dilution of χ_{lf} in these sediments and shows that evidence of increased erosion from this time exists.

The catchment was burnt by two bushfires, one in late 1904 and one a month later in early 1905. This has provided the opportunity to examine the effect of fire on the magnetic mineralogy of the catchment soils and to test whether such changes persist following transport and deposition. The charcoal from this fire is preserved in sediments at 0.98–0.94 m, which have been dated by ^{210}Pb methods to 1902–1920 AD. This shows again the rapidity of transport of eroded material from the catchment to the lake. Soils burnt in fires are generally considered to possess elevated concentrations of ferrimagnetic and superparamagnetic grains (Le Borgne, 1960; Dearing, 1999b). Burnt soils are therefore characterised by high values of χ_{lf}, SIRM and χ_{fd}, with this magnetic signature preserved once soils are eroded and deposited elsewhere (Le Borgne, 1960; Rummery *et al.*, 1979; Rummery, 1983; Oldfield and Clark, 1990, Dearing, 1999b). The sediments from this sub-unit appear to have no superparamagnetic grains (zero χ_{fd} values [Oldfield, 1999]), and the absence of spikes in χ_{lf} corresponding to the fired layers (0.98-0.94 m, Figure 6.4) seems to disagree with previous studies of burnt sediments. Part of the reason for this may lie with the low organic content of the soils at Tocal; for there is evidence that the burning of low-organic content soils has little impact on their magnetic mineralogy (Dearing *et al.*, 1996).

Although other explanations are possible, the consistent level of non-detrital P in the lake sediments (above 1.16 m depth, Figures 6.20 and 6.21) from this period probably records a stabilisation in animal populations in the immediate catchment after the early 1870s. After this time, a consistent flux of nutrients from agricultural activity and periods of drought and lowered water levels at Tocal led to the lake becoming more of a productive, seasonally inundated swamp than an open lake system. Evidence of this is clearly recorded in the visual log of this sub-unit, with the mottled sediments having greatly increased macrobotanical remains (Figures 6.3 and 6.4). The plant organic content of the sediments reaches its highest level between *c.* 1874 and *c.* 1912, with sediments deposited around 1900 composed of 29% organic matter. While it is possible that some of the plant organic content may have been eroded from the catchment, the sediments preserve a large volume of *in situ* plant remains, which probably accounts for the majority of this material (section 6.11.3).

Sediments in this unit are heavily mottled, containing extensive mats of roots and preserved stems and leaves, and possessing some evidence of early stage pedogenesis. This suggests that subaerial exposure of the lake floor may have taken place, while the mottling of this unit is consistent with a period of frequent wetting and drying during this period. The instrumental record shows clearly that the first decade of the 20th century was extremely dry (Figure 8.14) and, although conditions improved after *c.* 1907, wetter conditions did not recommence until *c.* 1912. The most substantial drying out of the lake probably took place at the end of this period, following the deposition of the majority of unit 2.2, as plant organic content is at its greatest near the top of the unit. Impacts on the catchment not just from agriculture, but also from extensive construction and deforestation, particularly in the first decade of the 20th century, were arguably severe. The combination of prolonged drought after 1900 AD, continued cattle operations, two bushfires, the removal of that natural vegetation that remained in the southeastern catchment and the construction of the North Coast Railway in the later half of the decade would have led to the severe degradation of soils and disturbance of slopes (section 4.6.3). Paradoxically, rates of soil erosion increased only marginally through this decade, from 160 to 183 t $km^{-1}a^{-1}$. While it is possible that erosion rates were limited by the depleted supply of readily erodible topsoil material, the temporary disconnection between the slope and lake basin may be related to drought in the early 20th century (section 8.3.6), which may have interrupted erosion and sediment transport at this time.

8.5.3 Unit 2.3: 0.94 m to 0.74 m (*c.* 1912 AD to *c.* 1936 AD)

Around 1912, successive waves of minerogenic material were stripped from the degraded slopes of the catchment and deposited in the lake basin (Figure 8.10), burying the swamp-like environment that had evolved in the pervious decades (see the marked change in sediments above 0.94 m in Figure 6.4, for example). This rise in sediment flux is clearly recorded in the sediment's magnetic mineralogy, with χ_{lf} values doubling by *c.* 1922 (Figure 6.7), while marked increases in many of the cations indicative of catchment erosion (Na, Mg, K and Ca) followed. Erosion rates in the first years of this period rocketed to over 800 t $km^{-2}a^{-1}$, suggesting that a limit on the supply of erodible material may not have been responsible for the low erosion rates recorded at the turn of the century (Figure 8.10). Two alternative explanations may account for the dramatic fluctuations in erosion during this period.

First, the severe drought conditions of the early 20th century limited the transport and deposition of soil in the lake basin. Support for this model has been detailed already for the first three decades of the 20th century (section 8.3.6). The *c.* 1912 spike in erosion coincides with years of high runoff at Tocal (Figure 8.12), which followed drought conditions that persisted from *c.* 1900 to 1907 (Figure 8.14). The volume of runoff in the years 1910–1913 was, however, equal in magnitude to that recorded in 1904 and 1908, when no corresponding peak in erosion was recorded. This would suggest that higher runoff alone may not be responsible for the fluctuations in erosion.

In contrast to the first theory, the second explanation suggests that catchment erosion was not interrupted during this period, and that fluctuations in the erosion record may relate to changes in the intensity or location of land use. Some support for the thesis of efficient sediment transport during this period comes from the charcoal record of the 1904–1905 fire, which was deposited rapidly in the basin after the event, while arguments developed early in this dissertation show that direct linkages are likely to have existed between the catchment and the lake basin through time (section 3.3). As detailed in section 8.3.7, the spike in erosion in *c.* 1912 coincides with the construction of the railway through the southwest corner of the Tocal catchment, which involved major excavations and disturbance of the catchment's soils. In the absence of any better explanations, the massive pulse of sediment into the lake in *c.* 1912 is thought to record the considerable disruption and erosion of soil following this event. If so, then this would explain the availability of soil for remobilisation during this period, following the previous decades of low rates of catchment erosion.

Although Tocal changed ownership in 1926, cattle grazing continued as the main land use thought this period (section 4.7.1). Its impact on the catchment's slopes is not thought to have waned until the 1930s, after which time the intensity of agricultural activity began to decline (section 4.7). This continual impact of stock activity is recorded in the peak levels of erosion that characterised this period, the maximum of over 930 t $km^{-1}a^{-1}$ being recorded at the beginning of the 1920s (Figure 8.10). There are some data that suggest that this erosive pulse followed years of drier conditions in the late 1920s, and coincided with the return of much wetter conditions capable of transporting soils disturbed in the previous decade into the lake (see '1922' in Figure 8.12). Although erosion was at its zenith during this period, the stratigraphy of core TCA9b suggests that only silt-sized material was deposited in the basin at this time (Figure 6.4). Although it is possible that coarser material may have been deposited proximally, deposition of coarse material in the central lake basin had occurred in earlier periods, demonstrating that such sorting processes are probably not responsible for the relatively fine texture of the sediments. Instead, the lack of coarser material may record a switch in sediment source to the fine silt-rich alluvial soils from the immediate slopes of the lake. Unfortunately, there is no information from the sediment provenance dataset on the origin of sediments from the 1920s.

Plant organic content dropped sharply after *c.* 1914 (Figure 6.5) as the massive increase in mineral sedimentation resulted in the burial of aquatic plants on the lake's bed. In addition, increased water levels in the lake probably had the effect of flushing the system, reducing its trophic status, and temporarily decreasing the lake's organic productivity. However, counteracting this in the following years was a marked increase in the flux of nutrients in the lake basin, with phosphorus levels in the lake reaching their maximum during this period, peaking around 1930 AD (0.78–0.76 m) (Figures 6.20 and 6.21).

Macroscopic crystals of vivianite ($Fe_3(PO_4)_2.8H_2O$) first appear in the Tocal record above 0.90 m (after *c.* 1921). Vivianite is an authigenic mineral formed when Fe in sediments is reduced to its ferrous form (Mackereth, 1966), and is indicative of deposition in an anoxic, reducing environment in the presence of excess P (Hernandez-Pacheco *et al.*, 1993; Olsson *et al.*, 1997). This suggest that after *c.* 1921 high rates of sedimentation and nutrient input led to highly eutrophic conditions in the lake, with anoxic, reducing conditions prevailing. This led to an increase in organic productivity. From the mid-1920s to the early 1930s, plant organic content remained stable at around 10%, but increased sharply after this time, reaching a peak at *c.* 1934 (0.78–0.76 m). Further evidence of a rise in productivity before *c.* 1934 comes from the spike in S at this depth, indicating an increase in lake eutrophication (section 6.11.2.5), probably exacerbated by the drought conditions that plagued the mid-1930s (Figure 8.14).

In the early to mid-1930s, the χ_{ARM}/χ_{lf} quotient increased by nearly 400%, reaching its highest value in the entire core at 0.76–0.74 m (Figure 6.13) (a similar rise is recorded in the χ_{ARM}/SIRM quotient). In conjunction with declines in χ_{lf} and SIRM at this depth, the magnetic evidence indicates a sharp drop in ferrimagnetic mineral deposition in the basin in the early 1930s. The most likely explanation for this is that rates of minerogenic erosion in the catchment decreased sharply after *c.* 1930 (Figure 8.10).

The mid-1930s to the early 1940s were dominated by drought conditions at Tocal (Figure 8.14), which may have interrupted the supply of sediment into the lake basin. Evidence of drought in sediments may be recorded by zero χ_{fd} values, as dissolution of magnetic minerals in the lake takes place (section 6.11.1.2). The presence of χ_{fd} grains in the sediments above 0.78 m (*c.* 1931) (Figure 6.10) suggests that these sediments contain ultrafine ferrimagnets (Maher *et al.*, 1999). The most likely source of these minerals is eroded catchment soil (these sediments have similar values to those of the pedogenic magnetites studied by Maher *et al.* [1999, pp. 41, 43]). The persistence of these minerals at this depth would suggest that drought conditions may not have impacted these sediments. However, an authigenic origin of these χ_{fd} grains from this period (possibly from greigite or pyrrhotite [iron sulphite]) cannot be ruled out entirely, particularly considering the accompanying increase in organic sulphur in the lake at this time.

An alternative explanation for the low erosion rates from the early 1930s is that they record some shift in the location or nature of land use. The 1930s saw the beginning of the end of agriculture at Tocal, with farming effectively stopping until the establishment of the college in the 1960s. No horses were kept at Tocal after 1927 and increasing numbers of cattle were sold from the property from the mid-1930s onwards (section 4.7.1). Agriculture is thought to have ground to a halt by about 1940. Declining agricultural activity fits well with the pattern of lower soil loss recorded through this period.

Sediments eroded from the catchment around 1930 AD (0.78–0.76 m depth in core TCA9b) are geochemically identical (within 95% confidence limits) to the catchment soil TS24s, located in the far southeastern corner of the catchment (Figure 5.6). However, for reasons outlined in section 6.8.5, little confidence is given to this sediment source ascription. Further doubt about the validity of this connection comes from the soil's location upslope of the North Coast Railway line, completed in 1912 AD, which effectively cut sediment transport from those slopes into the remainder of the catchment, and into the lake. An alternative hypothesis is there may have been a lag in the delivery of eroded material from the furthest corner of the catchment to the lake basin. Construction of the railway line in the early 1910s would have resulted in a great disturbance of soils in this area; however, the eroded material may not have been deposited into the lake until *c.* 1930 AD. This hypothesis is difficult to accept, as there is no evidence of significant lags in sediment delivery at other times in the catchment. The sediments from this period, therefore, have no match with any potential sediment sources in the catchment. However, this is probably a realistic reflection of sediment supply during this, and other periods, at Tocal. With soil erosion affecting much, if not all, areas of the catchment, the lake sediment represents a mixture of a range of chemically distinctive soil material. Trying to disentangle the relative chemical signatures of individual soil types eroded into the lagoon from a mixture is difficult at best, and may be impossible in some circumstances as contributing soils may no longer exist.

8.5.4 Unit 2.4: 0.74 m to 0.54 m (*c.* 1934 to *c.* 1957)

The period *c.* 1934 to *c.* 1957 spans those years when agricultural activity effectively halted, and two major bushfires damaged the Tocal property. The absence of any charcoal from the 1939 bushfire confirms that it had little impact in the catchment of Tocal Homestead Lagoon (section 4.7.2). The impact of the 1944 fire, however, is clear. Soil erosion rates rose from 325 to 529 t $km^{-2}a^{-1}$ in response to the fire, with macroscopic charcoal recorded at 0.64-0.60 m. The visual log of core TCA9b shows that low organic content silts were rapidly deposited in the lake basin, burying organic-rich sediments and aquatic plants that had developed in the productive conditions of the previous years (Figure 6.4).

The sharp drop in P levels during this period mimics the decline in stock at Tocal (Figure 6.22). As the impacts of agriculture during this period were much reduced, the catchment's erosional response is probably a record of the 1944 bushfire alone. Immediately after the fire, soil material predominantly from the T20 and T4w locations (the alluvial soils immediately around the lake and on the southeastern slope) (section 6.8.5) (Figure 5.6) was eroded into the basin. This corroborates the historical record, which suggests that the western and probably southwestern parts of the catchment were less affected by the fire (section 4.7.2). By *c.* 1955 the major sediment source had become soil type T18, located in the eastern part of the catchment.

Erosion rates immediately following the fire were probably attenuated to some extent by the drought conditions of the mid-1940s, as higher rates of soil loss were maintained until the end of the decade, despite a complete halt in agricultural activity after 1947. This phenomenon is recorded clearly in the magnetic record. Although a sharp increase in χ_{lf}, IRM, SIRM and HIRM dates from the 1944 fire, these values do not reach their maxima until 0.60–0.58 m (*c.* 1953). The increased intensity of erosion resulted in a coarser-grained magnetic mineralogy in the lake, with noticeable declines in χ_{ARM} and associated magnetic quotients.

Does this represent some time lag in erosion following the 1944 bushfire? The answer is probably not, considering the rapidity with which the charcoal from the event was delivered into the lake basin (Figure 6.4). It is more likely that the continued high rates of erosion and flux of ferrimagnetic minerals into the lake years after the fire record a period in which the soils at Tocal were restabilising. Bushfires oxidise soil organic matter and enhance water-repellent properties; the unprotected, hydrophobic surface soil aggregates are highly erodible until a ground cover is re-established (Prosser, 1990; Saynor *et al.*, 1994). The continued impact of the 1944 fire may have been assisted to some extent by increased runoff in the catchment from 1949 onwards (Figure 8.12), possibly delivering the last of the easily erodible material that was disturbed in the 1944 event.

The increased flux of sediment in the years after the 1944 bushfire appears to coincide with a marked decrease in lake productivity, recorded in the sediments as minima in LOI_{430} and in S concentrations at 0.66–0.64 m (*c.* 1946) (Figure 6.36). Increasingly eutrophic lake conditions (recorded in the LOI_{430} and S record) were experienced at Tocal after 1953, and reached a peak at the top of this unit (0.54–0.52 m [*c.* 1957]). There is some evidence that the increase in eutrophication may have been driven by higher levels of nutrient input into the basin (Figure 6.22). The presence of macroscopic vivianite in the sediments below 0.52 m also implies an environment of nutrient-rich anoxic lake waters. With the paddocks empty, greater vegetation cover across the slopes stabilised catchment soils (see, for example, Figure 4.18). As a result, a decrease in erosion and lake sedimentation was recorded, and there was a return to a prolonged period of quiescent conditions in the lake for the first time since European settlement.

8.5.5 Unit 2.5: 0.54 m to 0.28 m (*c.* 1959 to *c.* 1976)

The base of this unit records the Tocal environment immediately before the construction of the college, at the end of a 15 year period of negligible agricultural activity. Around 1960 AD, Tocal Homestead Lagoon was a highly productive swamp with low inputs of mineral soil from the well-vegetated catchment. Rates of catchment-wide soil loss were between 220 and 240 t $km^{-2}a^{-1}$. Average to above average runoff from the catchment probably meant lake levels were maintained at this time (Figure 8.12). Peak values of plant organic matter and S (essentially organic S [section 6.11.2.5]) reflect this: LOI_{430} values for this period are the highest recorded since the beginning of the 20th century. The construction of the college in 1964 involved extensive devegetation and earthworks (section 4.8.1), and rates of soil loss rapidly increased, doubling to 440 t $km^{-2}a^{-1}$ after this time. Increases in Fe, Mg, K and Zn resulted from the influx of soil material into the basin (Figure 6.22), while rises in χ_{lf}, $IRM_{100\ mT}$ and SIRM reflect the rapid influx of ferrimagnetic material. The reappearance of χ_{fd} grains in the sediment after construction of the college reflects the disturbance, erosion and deposition of topsoil into the lake.

Nearly as rapidly as these rates accelerated, they declined, remaining around 300 t $km^{-2}a^{-1}$ through the reminder of the 1960s. This is despite detailed evidence of rapidly increasing stock numbers at Tocal accompanying the college's growth through the decade (section 4.8, Figure 4.24). One explanation for the decline in erosion rates is that drought conditions during the 1960s may have interrupted the transport of soil to the lake. As shown in section 8.4.6, peaks and troughs in erosion during the 1960s and early 1970s fit very well with the pattern of floods and droughts. The correspondence of the two temporal trends, however, cannot be simply accepted as being a direct causative association. The period 1965–1966 was particularly dry at Tocal; possibly dry enough to limit erosion and sediment transport in the catchment. However, a longer and drier period recorded in the 1880s does not appear to have had a significant impact on erosion (Figure 8.14).

An alternative theory is that although agricultural activity was intensive, the Tocal catchment may have experienced some relief with stock relocated to other parts of the property beyond the catchment. The 1967 aerial photograph of Tocal shows extensive bare soil on the western slopes of the lake, and the incision of the usually inactive drainage lines that feed into the southern shore of the lake (Figure 8.7). By contrast, the 1958 air photograph shows these slopes to be much more stable, with no gullying. The first years of the college clearly had a significant impact on the catchment, and it may be that to ease pressure on the soils agricultural activity was transferred elsewhere. Such a response is practiced at Tocal today, but whether this was done in the recent past is unknown.

This thesis may be tested by examining the lake's geochemical record for evidence of a drop in stock numbers in the catchment. Sharp declines in both total P and non-detrital P (section 6.11.2.2) above 0.48 m mirror the decrease in soil erosion during the 1960s (Figure 6.21). This suggests that there was a reduction in the organo-phosphate input to the lake during this period, which is most plausibly explained by a reduction in stock numbers in the lake's catchment. Although it cannot be determined within any certainty, there is at least some geochemical evidence to support the stock-relocation explanation of lower erosion rates during the 1960s.

Aerial topdressing of superphosphate probably began in the Hunter valley around 1957. However, this appears to have contributed little to the geochemistry of Tocal Homestead Lagoon. Although there was an exponential increase in the volume of superphosphates delivered to land in New South Wales from the late 1950s to the mid-1960s (Table 4.9), P concentrations in the lake sediments of Tocal Homestead Lagoon decreased slightly through much of this period (Figure 6.21), which may reflect the absence of significant fertiliser application at Tocal.

In *c.* 1970, sediments deposited in the lake (0.42–0.40 m) were characterised by peak χ_{lf}, $IRM_{100\,mT}$, SIRM and HIRM values, associated with a sharp increase in catchment erosion to 455 t $km^{-2}a^{-1}$. Climatic processes appear to have played little role in this increase: no period of wetter conditions accompanied this sharp rise, while drought conditions had finished by the end of 1966 (Figure 8.14). By 1970, animal numbers at Tocal had increased by over 10 times by comparison with stock numbers at the time of the foundation of the college (Figure 4.24). Their disturbance of the catchment's soils may have been sufficient to accelerate catchment erosion (Figure 8.10). The mid- to late 1970s were largely characterised by a decline in sediment flux to the basin. This was recorded by both these magnetic proxies and the lake-sediment soil erosion record. This decrease cannot be explained by any decline in agricultural activity. Figure 8.15 clearly shows that the increase in stock numbers at Tocal continued unabated until 1981. Some explanation for the decrease erosion under these circumstances again comes from P record. Sharp declines in P ratios mimic the decrease in catchment erosion (Figure 6.35). Very similar patterns recorded in the mid-1960s were interpreted as the result of the relocation of hard-hoofed stock to other parts of the catchment, thus reducing both catchment erosion and nutrient input to the lake. It is likely that a similar pattern of land uses was responsible for the low erosion rates during the mid- to late 1970s. This suggests that the agricultural statistics from Tocal should not be interpreted without taking into account the size of the property and complexity of its land use history. Changes in stock numbers are not necessarily a direct measure of changes in agricultural activity in the catchment.

The sediments between 0.38 m and 0.36 m record a period of high LOI_{430} values and peaks in both S and P concentrations (Figure 6.22). The visual record of core TCA9b shows that plant stems and roots in growth position dominated the organic content of these sediments, suggesting that a relatively short-lived period of increased aquatic productivity occurred in the early 1970s. The sharp boundary with the overlying sediments may represent an unconformity. It is suggested that the sediments from 0.38–0.36 m underlying this unconformity may be contemporaneous with drier conditions. Such conditions would have the effect of reducing lake levels and encourage eutrophication of the lake, explaining the high plant organic matter and S and P peaks in these sediments. Drier conditions would also ultimately result in an episode of non-deposition, which would account for the unconformity overlying these sediments. The ^{210}Pb date of the base of the 0.38–0.36 m drought layer is 1971 (1964–1978 age range) (Table 7.7). During this period a dry year was recorded in 1973 (Figure 8.14). It is this drought that is most probably responsible for the organic rich sediments at 0.38–0.36 m,

A near identical event is recorded further up the sequence, in the sediments from 0.30–0.28 m, deposited in *c.* 1977 (Figure 6.4). These sediments have a similar coupling of peak LOI_{430}, P and S values. Drought conditions that persisted at Tocal from 1979–1980 (Figure 8.14) best explain the band of organic sediments preserved at 0.30–0.28 m.

8.5.6 Unit 2.6: 0.28 m to 0.00 m (*c.* 1978 to 1998 AD)

In the mid- to late 1980s (^{210}Pb date of 1987, 0.16 m depth in core TCA9b), a massive pulse of sediment was deposited in the lake (Figure 8.10). These sediments were characterised by the core's maximum χ_{lf} value, and spikes in χ_{ARM}, $IRM_{100\,mT}$, SIRM and HIRM (Figures 6.8 and 6.11). The magnetic evidence suggests that the soil-derived material consisted of relatively coarse ferrimagnetic minerals, while the visual record shows that the material consisted of silt and some fine sand (Figure 6.4). This event is clearly revealed as a spike in the lake's 2000 year catchment soil loss record, and follows drought-like conditions in the catchment through the early 1980s. After this time, soil loss from the catchment declined sharply, with the late 1980s and the 1990s having roughly half the soil loss of the mid-1980s. This decline seems to have little to do with total stock numbers at Tocal, which did not reach their maximum until the late 1980s (Figure 4.24).

Interestingly, the Tocal lake sediment sequence appears to have recorded the effect of the college's adoption of sustainable agricultural practices (Anon., 1997). Aerial photographs from the mid-1990s suggest that catchment stability was at its greatest since the construction of the college in the early 1960s. Drainage density in the catchment in 1996 was lower than for any other period since the 1950s (Figure 6.32). Across the Hunter valley, the total area affected by erosion decreased steadily from the mid-1940s to the 1970s (Figure 4.23). Findings from Tocal suggest that this trend was maintained and that soil loss in the 1990s was much less than that experienced during the second half of the 20th century, probably reflecting Australian society's increasing concern for the state of the environment.

The level of P in the lake sediments increased steadily after 1980, and reached its maximum value in the 2000 year record in the early 1990s (0.10–0.08 m, Figure 6.21). Peaks in S and LOI_{430} in the sediments deposited from this period probably record the combined impact of high nutrient levels and the drought conditions of the early 1990s, resulting in highly eutrophic conditions in the lake. The presence of algal blooms and water borne weeds in the lake through the 1990s (section 3.2) confirms the sedimentary evidence of lake eutrophication. It is also provides a modern example of those conditions experienced during earlier periods of organic-rich deposition. Interestingly, the eutrophic conditions of the 1990s did not result in any increase in plant growth on the lake floor during the period (Figure 6.4), as occurred during previous periods of eutrophic conditions at Tocal. This may relate to relatively-high lake levels throughout most of the period, and a corresponding decrease in sunlight to the lake bed.

Above 0.08 m, a large spike in magnetic parameters is recorded. This suggests a sharp rise in the concentration of ferrimagnetic minerals deposited in the mid-1990s, a spike similar in magnitude to that recorded around 1987 (0.16–0.14 m). Although a rise in catchment-wide erosion is recorded for this period (Figure 8.10), it is much lower than might be expected based on the magnetic properties of the sediments at this depth. Rather than just recording a phase of major sediment flux into the lake, these sediments may reflect an increase in the concentration of finer ferrimagnets eroded into the lake at this time. Some support for this comes from the sharp decline in the SIRM/χ_{lf} quotient above 0.10 m, which indicates a shift away from coarser grained ferrimagnets (Figure 6.13). Although no specific event in the catchment's recent history coincides with this record, the construction of roads, tracks, fences and buildings was common through the 1990s, and it is probable that this peak in finer grained ferrimagnetic mineral deposition in the lake is linked to an increase in topsoil erosion during this period.

8.6 A 2000 year record of environmental change at Tocal: summary

Variations in lake sedimentation and catchment denudation at Tocal have been the result of an often-complex interaction of human and climatic conditions during the last *c.* 2000 years. For much of the post-settlement record, sediment flux into the basin has been in phase with catchment land use, with excellent agreement between the historical record of land use at the site and major changes in soil loss. There is, however, some evidence of the influence of climatic processes (particularly extreme and prolonged drought) on catchment erosion: climatic perturbations have clearly amplified the impacts of agriculture at Tocal. The importance of high-magnitude droughts, and the apparent relative unimportance of decadal scale fluctuations in rainfall at Tocal before European settlement may reflect the impact of warmer conditions in the last century or so, which are considered to have been greater than any other period in the last 1000 years (Crowley, 2000; Osburn and Briffa, 2006). One important finding from the Tocal environmental record is that the dual processes of climate and land use cannot be considered independently when assessing their part in soil erosion. The individual roles of human activity, climate and natural phenomena such as fires have proven capable of being separated in this study, due in part to a high-resolution chronological control, but also to fortuitous historical circumstances.

Although climatic shifts appear to have had some part to play in the erosion story, they appear to have been insignificant at times of severe degradation in the catchment. The *c.* 50 year period of low minerogenic sedimentation in Tocal Homestead Lagoon following the first years of European settlement suggests that human activity and possibly sediment-supply limitation were the only controls during this period. This prolonged period of very stable and low sedimentation in the 19th century occurred despite a series of major floods and droughts that was experienced across central eastern Australia. Documented high-magnitude floods in the decades before European settlement at Tocal appear to have had little impact on catchment erosion, which reached the lowest levels recorded in the site's *c.* 2000 history. This finding is of major significance, as it suggests that recent climatic change may have been of relatively little consequence in catchments that have experienced extended periods of intense farming. A focus on historical land use, and an understanding of the erosion processes of soil may be much more important in interpreting long histories of catchment sediment yield.

The post-settlement history of fire at the site has shown that the burning of soils and the loss of humus and soil cover probably led to greatly enhanced erosion, at least over short timescales. However, fire alone has probably not been the cause of elevated erosion rates. Fires have often coincided with particularly dry conditions in the region, while it is has been land use and, possibly, runoff in the years after fire that exacerbated soil loss. Although the impact of fire on soil erosion at Tocal broadly confirms findings from other sites in central eastern Australia (Dodson and Thom, 1992; Johnson 2000), its effect in the catchment has been short lived, and longer term trends in agricultural activities, mainly the decadal scale cattle operations in the catchment, have resulted in greater soil loss.

The lake sediment sequence has provided evidence suggestive of Aboriginal activity at Tocal in the millennia before the arrival of Europeans, corroborating modern observations of their presence at the site in prehistory. Although the strongest evidence of pre-European human activity at Tocal dates from after 800 AD, Aboriginal occupation in the catchment may predate the beginning of the lake sediment record; regular firing of the landscape is recorded throughout the entire pre-European sequence (but for two brief hiatuses), which may have been the result of Aboriginal land use.

The environmental reconstruction has shown that, although conditions from *c.* 250 BC to 1820 AD appear stable in comparison with the post-settlement period at Tocal, significant fluctuations were recorded in the lake during this time. While the only other palaeoenvironmental records for this region show relatively stable conditions through this period, the Tocal record has produced new evidence that suggests that warmer and possibly drier conditions may have begun at Tocal sometime after 900 AD, and persisted well into the 15th century. These may correspond with the Medieval Warm Period identified around this time in many other parts of the world. These centuries at Tocal were characterised by low and stable sedimentation rates (the lowest recorded in the site's history), and the data suggest that a period of increased catchment stability, vegetation and, by extension, warmer conditions may have been the cause. However, more detailed dating and palaeoenvironmental data for the pre-European record is required before any certainty may be attached to this theory.

Butzer and Helgrin (2005) have proposed that the impacts of European settlers on the Australian environment were much less than many commentators believe, arguing that the impacts of Aboriginal land use and catastrophic climatic variations in Australia before colonisation led to a 'metastable' equilibrium, and a high resilience. Beyond the firing of the landscape, however, Aboriginal impacts on land degradation are insignificant compared to even the most rudimentary, small-scale modern agriculture that accompanied European settlement in Australia. This has been well-proven at Tocal, and probably holds true for similar temperate environments on the continent. While cycles of floods and droughts appear to have had little impact on soil loss prior to European settlement at Tocal, suggesting some level of 'resilience' of soils to shifts in climatic conditions, the immediacy and severity of erosion that accompanied the onset of modern agriculture in the 1820s demonstrates clearly how easily and rapidly this equilibrium could be disrupted. Lag times between the historical record and the lake sediment record at Tocal appear to have been insignificant, and in any case are less than the resolution of the record. This shows that 'non-linear land use limnological response relationships' (Tibby, 2003) cannot simply be assumed to be applicable *en masse* to the Australian environment.

While the resolution of environmental change at Tocal for the *c.* 2000 years before European settlement is coarse at present, it is clear that the long-term mean flux of sediment from catchment slopes was an order of magnitude less than that for the *c.* 180 years under European agriculture, while the immediate impact of European settlement was an increase in soil loss to levels 200 times greater than any climatic fluctuations or Aboriginal impacts had generated in the preceding 2000 years.

In the light of these findings, Butzer and Helgrin (2005) appeared to have grossly overestimated either the 'resilience' of the pre-European environment to disturbance, the impact of pre-European climatic fluctuations and Aboriginal impacts or probably all three. Alternatively or additionally, they have underestimated the impacts of European settlement on the Australian environment. This *c.* 2000 year high-resolution record of catchment-wide denudation from eastern Australia lends strong support to the argument that the arrival of Europeans, and the onset of modern agriculture, led to the greatest environmental change recorded in the late Holocene period.

Chapter Nine

Conclusions

9.1 Introduction

The main aim of this research has been to develop a long-term, well-dated record of environmental change from the sediments of Tocal Homestead Lagoon. In doing so, the impact of human activity on catchment denudation through the late Holocene may be assessed. Particular attention has been given to the accurate and precise dating of the point of European contact in the sedimentary record and, in doing so, significant advances in the dating of recent sediments in the Australian environment have been made. With the record of first European disturbance known with confidence, it has been possible to obtain specific information on the timing and nature of European environmental impacts at this site. In addition, insights into the state of the pre-impact environment during the last *c.* 2000 years have been gained. These include possible evidence of Aboriginal land use in the catchment and indications of periods of contrasting climatic conditions during the late Holocene in eastern Australia. Finally, findings from this research have implications for ongoing debates on geomorphic processes, erosion and human impacts on catchments.

9.2 Soil erosion, land use and climatic fluctuations

Through the use of multi-method dating of the Tocal sediments, the depth in the lake record corresponding to the point of European settlement in the catchment is known with confidence. Using this as a basis and employing the well-established basin-wide magnetostratigraphy, rates of lake-wide sediment deposition, and hence estimates of catchment-wide minerogenic soil loss may be made for the pre- and post-settlement periods. The mean rate of catchment-wide soil loss for the *c.* 2000 years before the arrival of Europeans was 18 t $km^{-2}a^{-1}$. This may be contrasted with the mean rate of erosion experienced under European land use of 228 t $km^{-2}a^{-1}$, an increase of more than an order of magnitude. Although this mean rate for the post-settlement period is comparable to modern rates of catchment sediment yield determined throughout eastern Australia, it masks periods of much greater denudation. The arrival of Europeans at Tocal is clearly recorded in the lake sediments as a pulse of topsoil-dominated material, which marked a spectacular rise in erosion to 638 t $km^{-2}a^{-1}$, more than 200 times higher than the rate at which soils were being eroded in the preceding four centuries. With only parts of the lake's catchment potentially under agriculture in the first years of settlement, erosion rates locally may have been in excess of 1200 t $km^{-2}a^{-1}$. Truncated A-horizons and compacted soils in the catchment today provide modern records of this past denudation. The catchment's response to the arrival of Europeans, the clearing of land for farming and the impact of thousands of hard-hoofed animals in the catchment was both swift and intense. In contrast, the pollen record for this period shows little variation.

Although the initial shock of European impact resulted in a dramatic increase in erosion, the greatest rates of catchment-wide soil loss at Tocal took place during periods of intense cattle ranching at the beginning of the 20th century. Earlier, when agriculture at the site was crop-dominated, the catchment appears to have been (relatively) more stable and erosion less severe. Lower rates of erosion have been recorded in the last decade at Tocal, which may represent the impact of more sustainable farming practices. This pattern is somewhat different to that recorded in other long-term sedimentation records in eastern Australia, which typically show that the greatest erosion rates immediately followed European settlement. This suggests that the interaction of land use and catchment erosion at Tocal is quite complex and has been complicated by shifts in land use, property size, catchment geomorphology and possibly by gaps in the historical record.

Decadal-scale climatic variations appear to have had little direct influence on the history of erosion, although there is good evidence that extreme climatic conditions may have been responsible for amplifying the already severe impacts of intensive land use. This may be in part due to the dominance of land use in controlling erosion in the historic period. Observed cycles of drought and flood in the region in the decades before European settlement in *c.* 1820 appear to have had no impact on soil erosion either. It appears that the natural environment may have had some level of resilience to short-term climatic fluctuations; consistent catchment stability being much easier to maintain under the protection of denser tree canopies and soil vegetation cover. If some level of equilibrium existed in the pre-European environment, as is suggested by Butzer and Helgren (2005), then it was both easily and rapidly disturbed and any resilience such systems may have had to minor climatic fluctuations appears to have been incapable of coping with the environmental changes that accompanied the arrival of Europeans.

9.3 Aboriginal occupation and land use

The Tocal Homestead Lagoon record shows that burning of the immediate catchment has taken place regularly since *c.* 150 BC. This is in agreement with every other charcoal record in the region, which show that regular firing of the environment has occurred during the last several millennia. However, the charcoal record alone cannot be simply interpreted as evidence of Aboriginal firing and land use at Tocal. More explicit evidence of this comes from the record of P (and to a lesser extent, Ca, Mg and Na) in the lake sediments, which displays a distinct increase above background levels from *c.* 800 AD, roughly 1000 years before European settlement in the region. By the time the first European had entered the catchment, P levels had already doubled above the site's pre-800 AD level. In the absence of any other plausible explanation for this increase, it has been cautiously interpreted as a record of human activity within the Tocal catchment. The timing of this pre-European disturbance is relatively recent compared to the chronology of Aboriginal occupation in the region, where most sites date to *c.* 2000 cal. BP. Although the arrival of Aboriginal people at Tocal 1000 years later is possible, an alternative hypothesis is that the date of *c.* 800 AD marks a new, more substantial phase of occupation at the site and that more sporadic occupation predated this, as recorded in the macroscopic charcoal record. More detailed charcoal and palynological analysis in the future may help to clarify the evidence of pre-European human activity.

9.4 Late Holocene environmental conditions

In providing much needed data on pre-impact environmental conditions at Tocal, important information on late Holocene environmental change has been uncovered. Changes in plant organic content and other sedimentological indicators around *c.* 200 BC were interpreted as representing a period of warmer and drier conditions at Tocal. At this time the lake's water level was probably low and fluctuating. Based on a linear interpolation of ^{14}C dates, it is estimated that this period finished in *c.* 400 AD. A dramatic shift in sediment magnetic properties is recorded sometime after *c.* 900 AD. A prolonged period of very low ferrimagnetic mineral deposition, low lake sedimentation and decline in fern species probably continued until *c.* 1200 AD. This may represent a period of increased catchment stability at Tocal, probably driven by a prolonged period of warmer conditions. The timing of this feature corresponds well with evidence of a Medieval Warm Period in the temperate mid-latitudes of the North island of New Zealand, with climatic simulations for the southern hemisphere and with the Tasmanian tree-ring record. There is currently little evidence of a Medieval Warm Period in mainland Australia. However, this may be explained by the near absence of dated palaeoenvironmental records with detailed information on the last 1000 years from the region. More high-resolution records of the last millennium are needed before climatic events such as the Medieval Warm Period and the Little Ice Age may be identified with any confidence in Australia. Until then, the Tocal record may provide some of the only evidence of warmer conditions coeval with the Medieval Warm Period in mainland Australia.

9.5 Implications for erosional processes

9.5.1 Introduction

Although this thesis has been concerned mainly with reconstructing the long-term response of a catchment to European land use, it has, also, revealed information of relevance to our understanding of geomorphic processes and catchment denudation since European settlement. These findings are discussed below.

9.5.2 Rapidity of environmental response

The geomorphology of the Tocal site, with its small catchment, short and relatively steep slopes and no significant sediment traps has promoted the rapid and direct delivery of eroded soil into the lake basin. As a result, lags between events in the catchment and their imprint on the lake sediment record appear to have been negligible, and almost certainly lie within the resolution of both the sediment sampling interval and the chronology. The first spike in sediment flux to the lake dates from *c.* 1822, coincident with the date of European settlement at Tocal. Later in the record, discrete layers of charcoal from documented bushfires in the 1900s and 1940s are preserved in ^{210}Pb dated sediments of the same age, again showing that direct links have existed between the catchment and the lake through the site's history.

9.5.3 Sediment supply limitation

In the Australian environment, which typically has thin, slow forming soils, rates of erosion may be controlled by sediment-supply limitation (Gale and Haworth, 2005). A striking 50 year period of sustained and low erosion which began only years after European settlement at Tocal distinctly resembles the profile of a sediment-supply limited system. As this period follows the first decades of severe erosion at Tocal, it is possible that the supply of easily-eroded topsoil may have been rapidly depleted and that the rate of catchment-wide erosion was limited. However, this trend may also be explained by the relocation of farming to areas beyond the immediate catchment of the lake. Both hypotheses are defensible, and both may in fact have been responsible for the 50 year period of low erosion. Degradation of the catchment in the first decade of agriculture may have prompted the relocation of stock and crops to undisturbed parts of the Tocal property, providing new food sources for stock and good soil for crops.

9.5.4 Erosion processes

The spatial pattern of sediment accumulation in Tocal Homestead Lagoon suggests that catchment erosion since European settlement has been dominated primarily by processes of hillslope erosion. Immediately above the horizon of European contact in the sedimentary record, thick accumulations of sediment were deposited in the central and southeast parts of the lake. This has provided some insight into the location of agricultural activity in the early years of settlement in the catchment, supplementing gaps in the historical record. In contrast, the sediment input to the basin via the drainage system appears to have been negligible. This, combined with aerial photographic evidence, suggests that extension of the drainage network or gullying has not been significant within the Tocal Homestead Lagoon catchment since European settlement.

This finding is in contrast to the work of Wasson *et al.* (1988) and Olley and Wasson (2003) who have proposed that erosion on the Southern Tablelands of New South Wales in the post-settlement period was dominated by channel incision. One explanation for the contrasting erosional histories may be related to climate; the tablelands are typically drier and milder than the temperate humid coastal areas of central eastern Australia. Soil loss by raindrop, wash or rill erosion is thought to take place mainly when soils are saturated (Selby, 1993). Sites in the Southern Tablelands region may receive as little as half the annual rainfall that sites in coastal eastern Australia receive, while Olley *et al.* (2003) note that the rainfall is distributed relatively evenly throughout the year. Catchments with higher rainfall such as Tocal, therefore, may experience conditions suitable for hillslope erosion much more frequently than those studied on the Southern Tablelands.

However, studies in similar environments by Gale and Haworth (2005) have found no evidence to suggest that gullying has ever taken place in the catchment of Little Llangothlin Lagoon on the Northern Tablelands of New South Wales. Instead, Gale and Haworth (2005) proposed that contrasting parent materials and soils may explain the differences in dissection and gullying in the landscape. At Tocal, however, the clastic bedrock materials are similar to those underlying much of the Southern Tablelands, and similar soil types may be found in both regions.

An alternative explanation may be presented to account for the range of erosional histories. Differences between slope and channel-erosion dominated systems in eastern Australia may be explained by differences in temporal and spatial scales and soil types. First, large-scale erosion surveys of the Hunter valley region suggest that, although erosion was mainly achieved by gullying in the 1950s and 1960s, sheet erosion has been the dominant form of erosion since the 1970s. This regional trend is similar to that recorded at Tocal (except that channel incision recorded in the 1950s–1960s was relatively minor). Contraction of the already small drainage network was recorded from the 1950s to today, probably in response to increases in the dominance of slope processes at the site. This suggests a strong temporal trend in erosional processes, and means that differences in the erosion mechanisms between catchments may be time dependant.

A final explanation for differences in erosion patterns may relate to the effects of soils and substrates. Gullying is most common on unconsolidated or mechanically weak materials (Stocking, 1980). The post-settlement channel incision on the Southern Tablelands discussed by Wasson *et al.* (1998), for example, was into either the alluvium substrate of valley floors, or colluvial deposits composed of A-horizon material eroded from upslope. In contrast, only a small fraction of the Tocal catchment is underlain by alluvium (the low relief slopes immediately surrounding the lake basin). In any case, much of the soil in the Tocal catchment is characterised by clay-rich B-horizons, which are comparatively far more resistant to incision than the soils and substrates that have been gullied in the Southern Tablelands region.

9.6 Implications for the management of soil erosion

The history of soil erosion at Tocal has provided insights into the behaviour of the catchment over long time spans. With the knowledge of pre-impact environmental conditions, sediment provenance, hillslope processes and land use impacts, some comments on soil conservation measures may be made. First, the years of greatest catchment denudation at Tocal were in the past and erosion today is, in contrast, nowhere near as problematic. Present day rates of catchment-wide minerogenic soil loss are in the order of 300 t $km^{-2}a^{-1}$, which is still an order of magnitude greater than estimates of soil replenishment by pedogenesis. Under such conditions, careful monitoring of stock densities (shown to be the cause of greatest erosion at Tocal), and the regeneration of paddocks showing signs of impact will aid in keeping farming environmentally sustainable. Secondly, soil loss from the catchment has been dominated by slope processes since European settlement, and possibly before this. Soil conservation measures, therefore, must be targeted to these processes.

Erosion banks have been constructed on the southern slopes of the lake in recent years. Although they have some ability to prevent erosion by the interruption of overland flow, such benefits are confined to those soils below the barriers. Given the dominance of slope erosion in the catchment, they are however an appropriate means of controlling sediment flux from upper slopes in the catchment. In contrast, gullying in the catchment appears to have been of little importance in the last *c*. 175 years.

A focus on channel-bank revegetation and stabilisation, championed by Wasson *et al.* (1998) as the best remedy for accelerated erosion, would probably not be repaid in a reduction in sediment yield in this small catchment. Furthermore, the results from this site, when compared with similar research from the region, suggest that a one-size–fits-all approach to soil conservation measures is not applicable, and that the characteristics of individual catchments should be considered before the limited resources available for soil conservation are eroded away.

9.7 Future research

Preliminary study of a longer sequence of sediment from Tocal Homestead Lagoon than that discussed here suggests that the site may have preserved an environmental record for most of the Holocene. Given the potential shown here for the preservation of information on sub-millennial environmental change and human activity, development of a longer Holocene record of similar detail would contribute greatly to our understanding of Holocene environmental change and human activity in eastern Australia.

The magnetic characterisation of lake sediments in this study has proven to be a valuable source of information on environmental change. Yet these methods have only rarely been used in Australia. Tocal Homestead Lagoon is the only site in Australia yet to have produced a detailed environmental magnetism record. Even so, further magnetic characterisation of these sediments has the potential to provide even more specific palaeoenvironmental data, through the use of detailed remanence measurements and temperature-dependant measurements.

Reference has been made in this research to the preliminary pollen analysis of core TCA9b undertaken by Dr Dan Penny (Cook *et al.*, in press). The robust, detailed chronology and palaeoenvironmental record established here will provide an excellent opportunity for comparison with the late Holocene pollen record. Future work on the fossil pollen record from Tocal Homestead Lagoon has the potential, therefore, to contribute significantly to current debates in Australia about the nature of vegetation response to European settlement, and the state of the pre-impact environment.

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Appendix 1

Volume-specific magnetic susceptibility (VSMS) measurements made on lake sediment cores from Tocal Homestead Lagoon, Paterson, eastern Australia. VSMS is in units of 10^{-5} SI units, 'VSMS 1' and 'VSMS 2' are the first and second measurements made on each core. Measures of error represent 95% confidence limits of the measurements.

Core TCA1					**Core TCA2**				
Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error	Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.04	86	83	84.5	6.5	0.04	249	246	247.5	6.5
0.11	123	120	121.5	6.5	0.11	309	309	309.0	0.0
0.18	158	158	158.0	0.0	0.18	202	206	203.8	9.7
0.25	197	199	198.0	4.3	0.25	148	150	149.0	4.3
0.32	246	246	246.0	0.0	0.32	144	143	143.5	2.2
0.39	238	234	236.0	8.6	0.40	150	150	149.8	1.1
0.46	201	199	200.0	4.3	0.47	149	148	148.3	1.1
0.53	186	184	185.0	4.3	0.54	150	149	149.5	2.2
0.60	180	178	179.0	4.3	0.61	156	155	155.5	2.2
0.67	171	170	170.5	2.2	0.68	155	154	154.5	2.2
0.75	162	161	161.5	2.2	0.75	121	123	121.8	3.2
0.82	160	159	159.5	2.2	0.83	97	97	97.0	0.0
0.89	157	157	157.0	0.0	0.90	84.5	85	84.8	1.1
0.96	161	160	160.5	2.2	0.97	89.5	90	89.8	1.1
1.03	148	147	147.5	2.2	1.04	56	58	57.0	4.3
1.10	115	116	115.5	2.2	1.11	31	31	31.0	0.0
1.17	91	89	90.0	4.3	1.19	23	24	23.5	2.2
1.24	66	65	65.5	2.2	1.26	22	23	22.5	2.2
1.31	53	52	52.5	2.2	1.33	23	24	23.5	2.2
1.38	52	53	52.5	2.2	1.40	22	22	22.0	0.0
1.46	41	40	40.5	2.2	1.47	23	22	22.5	2.2
1.53	30	30	30.0	0.0	1.55	20	20	20.0	0.0
1.60	25	25	25.0	0.0	1.62	18	18	18.0	0.0
1.67	23	22	22.5	2.2	1.69	17	17	17.0	0.0
1.74	23	22	22.5	2.2	1.76	17	17	17.0	0.0
1.81	21	20	20.5	2.2	1.83	17	17	17.0	0.0
1.88	21	20	20.5	2.2	1.90	16	17	16.5	2.2
1.95	20	20	20.0	0.0	1.98	17	17	17.0	0.0
					2.05	18	18	18.0	0.0
					2.12	18	19	18.5	2.2
					2.19	17.5	18	17.8	1.1
					2.26	18	17.5	17.8	1.1
					2.34	17.5	18	17.8	1.1
Core TCA3					**Core TCA4**				
Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error	Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.03	87	87	87.0	0.0	0.02	49	49	49.0	0.0
0.09	111	113	111.8	3.2	0.07	51	51	51.0	0.0
0.15	161	161	161.0	0.0	0.12	52	53	52.5	2.2
0.21	222	221	221.5	2.2	0.17	75	74	74.5	2.2
0.27	216	214	215.0	4.3	0.21	142	140	141.0	4.3
0.32	126	127	126.5	2.2	0.26	194	192	193.0	4.3
0.38	102	102	102.0	0.0	0.31	245	243	244.0	4.3
0.44	105	105	104.8	1.1	0.36	232	235	233.5	6.5
0.50	110	111	110.5	2.2	0.40	132	131	131.5	2.2
0.56	119	119	119.0	0.0	0.45	98	98	98.0	0.0
0.62	118	118	118.0	0.0	0.50	107	106	106.5	2.2
0.68	97	95	96.0	4.3	0.55	120	120	120.0	0.0
0.74	70.5	71	70.8	1.1	0.59	112	113	112.5	2.2
0.80	61	61	61.0	0.0	0.64	122	122	122.0	0.0
0.85	48	48	48.0	0.0	0.69	140	140	140.0	0.0
0.91	33	33	33.0	0.0	0.74	168	169	168.3	1.1
0.97	28	28	28.0	0.0	0.78	144	144	144.0	0.0
1.03	24	24	24.0	0.0	0.83	117	118	117.5	2.2
1.09	18	18	18.0	0.0	0.88	73	75	74.0	4.3

1.15	15	14.5	14.8	1.1	0.93	54	55	54.5	2.2
1.21	13	13	13.0	0.0	0.97	53	54	53.5	2.2
1.27	12	13	12.5	2.2	1.02	56	57	56.5	2.2
1.33	12	12	12.0	0.0	1.07	47	47.5	47.3	1.1
1.38	12	12	12.0	0.0	1.12	38	39	38.5	2.2
1.44	12	12	12.0	0.0	1.16	47	47	47.0	0.0
1.50	13	13	13.0	0.0	1.21	63	63.5	63.3	1.1
1.56	12.5	12.5	12.5	0.0	1.26	47	49	48.0	4.3
1.62	13	13	13.0	0.0	1.31	30	31	30.5	2.2
1.68	14	13.5	13.8	1.1	1.35	30	31	30.5	2.2
1.74	14.5	15	14.8	1.1	1.40	32	32	32.0	0.0
1.80	14	14	14.0	0.0	1.45	32	33	32.5	2.2
1.86	12.5	12	12.3	1.1	1.50	26	26	26.0	0.0
					1.54	20	20	20.0	0.0
					1.59	15	15	15.0	0.0
					1.66	12	13	12.5	2.2
					1.69	12	13	12.5	2.2
					1.73	12	13	12.5	2.2
					1.78	12.5	12.5	12.5	0.0
					1.83	12	13	12.5	2.2
					1.88	12.5	13.5	13.0	2.2
					1.92	13	14	13.5	2.2
					1.97	14	14	14.0	0.0
					2.02	13	14	13.5	2.2
					2.07	11	12	11.5	2.2

Core TCA5

Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.03	41	41	41.0	0.0
0.09	50	50	50.0	0.0
0.14	51.5	52	51.8	1.1
0.20	54	52.5	53.3	3.2
0.26	66	67	66.5	2.2
0.31	108	108	108.0	0.0
0.37	159	160	159.0	2.2
0.43	196	197	196.5	2.2
0.49	158	157	157.0	2.2
0.54	99	99	99.0	0.0
0.60	85	86	85.5	2.2
0.66	93	93	93.0	0.0
0.71	115	115	115.0	0.0
0.77	121	122	121.5	2.2
0.83	117	118	117.5	2.2
0.89	106	107	106.5	2.2
0.94	96	96	96.0	0.0
1.00	125	125	125.0	0.0
1.06	113	113	113.0	0.0
1.11	105	106	105.5	2.2
1.17	81	81	81.0	0.0
1.23	57	57	57.0	0.0
1.29	70	69	69.5	2.2
1.34	93	93	93.0	0.0
1.40	83.5	83.5	83.5	0.0
1.46	90.5	91	90.8	1.1
1.51	67	67	67.0	0.0
1.57	46	47	46.5	2.2
1.63	39	40	39.5	2.2
1.69	40	40	40.0	0.0
1.74	32	33	32.5	2.2
1.80	26	26	26.0	0.0
1.86	22	23	22.5	2.2
1.91	19	20	19.5	2.2
1.97	18	18	18.0	0.0
2.03	15	16	15.5	2.2
2.09	14	15	14.5	2.2
2.14	15	16	15.5	2.2
2.20	18	18	18.0	0.0
2.26	20	20	20.0	0.0
2.31	20	21	20.5	2.2
2.37	21	22	21.5	2.2
2.43	21	22	21.5	2.2

Core TCA6

Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.03	35	32	33.5	6.5
0.10	46	43	44.5	6.5
0.16	54	51	52.5	6.5
0.23	65.5	63	64.3	5.4
0.30	97	95	96.0	4.3
0.36	155	153	154.0	4.3
0.43	149	147	147.8	3.2
0.49	98.5	98	98.3	1.1
0.56	80	78	79.0	4.3
0.63	84	82	83.0	4.3
0.69	90	88	89.0	4.3
0.76	114	112	112.8	3.2
0.82	112	111	111.5	2.2
0.89	98	96	97.0	4.3
0.96	78	76	77.0	4.3
1.02	61	59	60.0	4.3
1.09	60	58	59.0	4.3
1.15	81	80	80.5	2.2
1.22	82	79	80.5	6.5
1.28	63	60	61.5	6.5
1.35	54	49	51.5	10.8
1.42	54	49	51.5	10.8
1.48	61	59	60.0	4.3
1.55	71	69	70.0	4.3
1.61	58	57	57.5	2.2
1.68	40	40	40.0	0.0
1.75	34	32	33.0	4.3
1.81	34	32	33.0	4.3
1.88	35	33	34.0	4.3
1.94	31	29	30.0	4.3
2.01	23	21	22.0	4.3
2.08	17	15	16.0	4.3
2.14	15	13	14.0	4.3
2.21	15	13	14.0	4.3
2.27	16	14	15.0	4.3
2.34	18	16	17.0	4.3
2.40	20	17.5	18.8	5.4
2.47	20	19	19.5	2.2
2.54	22	20	21.0	4.3
2.60	22	20	21.0	4.3
2.67	21	19	20.0	4.3
2.73	20	16	18.0	8.6

Core TCA7					Core TCA8				
Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error	Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.02	24	24	24.0	0.0	0.03	28	27	27.5	0.7
0.07	37	38	37.5	2.2	0.09	35	35	35.0	0.0
0.11	41	42	41.5	2.2	0.15	39	38	38.5	0.7
0.15	43	45	44.0	4.3	0.20	40	40	40.0	0.0
0.20	49	51	50.0	4.3	0.26	45	45	45.0	0.0
0.24	73	74	73.5	2.2	0.32	62	63	62.5	0.7
0.29	126	125	125.5	2.2	0.38	105	108	106.5	2.1
0.33	143	146	144.5	6.5	0.44	142	142	142.0	0.0
0.38	96	98	97.0	4.3	0.49	96	98	97.0	1.4
0.42	74	75	74.5	2.2	0.55	70	70	70.0	0.0
0.46	73	75	74.0	4.3	0.61	73	73	73.0	0.0
0.51	95	96	95.5	2.2	0.67	73	73	73.0	0.0
0.55	115	116	115.5	2.2	0.73	76	75	75.5	0.7
0.60	104	105	104.3	3.2	0.79	84	84	84.0	0.0
0.64	97	98	97.5	2.2	0.81	94	94	94.0	0.0
0.69	69	69	69.0	0.0	0.90	98	98	98.0	0.0
0.73	56	57	56.5	2.2	0.96	66	65	65.5	0.7
0.77	53	55	54.0	4.3	1.02	53	53	53.0	0.0
0.82	59	61	60.0	4.3	1.08	49.5	50	49.8	0.4
0.86	67	69	68.0	4.3	1.13	54	55	54.5	0.7
0.91	57	58	57.5	2.2	1.19	57	56	56.5	0.7
0.95	39	41	40.0	4.3	1.25	71	71	71.0	0.0
0.99	36	37	36.5	2.2	1.31	73	73	73.0	0.0
1.04	34	35	34.5	2.2	1.37	60	59	59.5	0.7
1.08	32	33	32.5	2.2	1.43	43	43	43.0	0.0
1.13	46	47	46.5	2.2	1.48	41	41	41.0	0.0
1.17	65	67	66.0	4.3	1.54	44	44	44.0	0.0
1.22	67	69	68.0	4.3	1.60	53	53	53.0	0.0
1.26	50	52	51.0	4.3	1.66	58.5	59	58.8	0.4
1.30	34	36	35.0	4.3	1.72	59	59	59.0	0.0
1.35	31	33	32.0	4.3	1.77	57	57	57.0	0.0
1.39	28	30	29.0	4.3	1.83	51	51	51.0	0.0
1.44	24	25	24.5	2.2	1.89	48	47	47.5	0.7
1.48	26	28	27.0	4.3	1.95	35	35	35.0	0.0
1.53	31	32	31.5	2.2	2.01	28	28	28.0	0.0
1.57	34	36	35.0	4.3	2.07	25	25	25.0	0.0
1.61	35	36	35.5	2.2	2.12	25	25	25.0	0.0
1.66	24	26	25.0	4.3	2.18	26	26	26.0	0.0
1.70	18	19	18.5	2.2	2.24	24	24	24.0	0.0
1.75	13	15	14.0	4.3	2.30	20	20	20.0	0.0
1.79	11	13	12.0	4.3	2.36	14	14	14.0	0.0
1.84	11	13	12.0	4.3	2.41	12	12	12.0	0.0
1.88	11	13	12.0	4.3	2.47	11	12	11.5	0.7
1.92	11	13	12.0	4.3	2.53	12	12	12.0	0.0
1.97	10	12	11.0	4.3	2.59	13	13	13.0	0.0
2.01	10	11	10.5	2.2	2.65	14	14	14.0	0.0
2.06	10	12	11.0	4.3	2.71	16	16	16.0	0.0
2.10	10	12	11.0	4.3	2.76	15	15	15.0	0.0
2.14	11	12	11.5	2.2	2.82	12	12	12.0	0.0
2.19	11	13	12.0	4.3					
2.23	11	13	12.0	4.3					
2.28	11	13	12.0	4.3					
2.32	10	12	11.0	4.3					
2.37	10	12	11.0	4.3					
2.41	10	12	11.0	4.3					
2.45	10	12	11.0	4.3					
2.50	11	13	12.0	4.3					
2.54	11	13	12.0	4.3					
2.59	13	14	13.5	2.2					
2.63	15	16	15.5	2.2					
2.68	15	18	16.5	6.5					
2.72	17	18	17.5	2.2					
2.76	16	18	17.0	4.3					
2.81	13	15	14.0	4.3					
Core TCA9a					Core TCB1				
Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error	Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.03	23	24	23.5	2.2	0.01	14	13	14.0	2.2
0.10	29	30	29.5	2.2	0.04	33	34	33.0	2.2

0.16	31	31	31.0	0.0	0.07	66	66	66.0	0.0
0.23	31	32	31.5	2.2	0.10	128	126	128.0	4.3
0.29	33	34	33.5	2.2	0.12	170	170	170.0	0.0
0.36	47.5	49	48.3	3.2	0.15	206	204	206.0	4.3
0.42	91	90	90.5	2.2	0.18	250	250	250.0	0.0
0.49	126	126	126.0	0.0	0.21	291	291	291.0	0.0
0.55	111	110	110.5	2.2	0.23	294	296	294.0	4.3
0.62	76	76	76.0	0.0	0.26	287	290	287.0	6.5
0.68	65	65	65.0	0.0	0.29	312	315	312.0	6.5
0.75	62	63	62.5	2.2	0.32	394	394	394.0	0.0
0.81	43	43	43.0	0.0	0.34	501	504	501.0	6.5
0.88	46	46	46.0	0.0	0.37	572	581	572.0	19.4
0.94	48	48	48.0	0.0	0.40	627	635	627.0	17.2
1.01	62	61	61.5	2.2	0.42	749	751	749.0	4.3
1.07	65	65	65.0	0.0	0.45	964	963	964.0	2.2
1.14	43	44	43.5	2.2	0.48	1234	1247	1234.0	28.0
1.20	40	40	40.0	0.0	0.51	1652	1663	1652.0	23.7
1.27	43	44	43.5	2.2	0.53	2055	2055	2055.0	0.0
1.33	47	47.5	47.3	1.1	0.56	2043	2034	2042.5	18.3
1.40	45	45	45.0	0.0	0.59	1639	1634	1638.5	9.7
1.46	42	42	42.0	0.0	0.62	1087	1090	1087.0	6.5
1.53	38	38	38.0	0.0	0.64	925	927	925.0	4.3
1.59	39	39	39.0	0.0	0.67	1083	1089	1083.0	12.9
1.66	44	45	44.5	2.2	0.70	1226	1230	1226.0	8.6
1.72	35	36	35.5	2.2	0.73	941.5	934	941.5	16.1
1.79	32	32	32.0	0.0	0.75	466.5	462	466.5	9.7
1.85	32	32	32.0	0.0	0.78	230	234	230.0	8.6
1.92	31	32	31.5	2.2	0.81	109.5	116	109.5	14.0
1.98	35	35	35.0	0.0	0.84	61	65	61.0	8.6
2.05	47	47	47.0	0.0	0.86	45	45	45.0	0.0
2.11	47	47	47.0	0.0	0.89	35	36	35.0	2.2
2.18	44	45	44.5	2.2	0.92	30	31	30.0	2.2
2.24	41	41	41.0	0.0	0.95	26	27	26.0	2.2
2.31	38	38	38.0	0.0	0.97	24	24	24.0	0.0
2.37	42	42	42.0	0.0	1.00	19	19	19.0	0.0
2.44	34	35	34.5	2.2					
2.50	19	19	19.0	0.0					
2.57	13	13	13.0	0.0					
2.63	13	13	13.0	0.0					
2.70	13	13	13.0	0.0					
2.76	13	13	13.0	0.0					
2.83	13	13	13.0	0.0					
2.89	11	11	11.0	0.0					
2.96	9	10	9.5	2.2					
3.02	12	12	12.0	0.0					
3.09	13	13	13.0	0.0					
Core TCB2					**Core TCB3**				
Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error	Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.01	51	51	51.0	0.0	0.02	62	62	62.0	0.0
0.04	73	74	73.0	2.2	0.07	86	87	86.5	0.7
0.07	74.5	75	74.5	1.1	0.11	118	118	118.0	0.0
0.10	77	77	77.0	0.0	0.15	180	182	181.0	1.4
0.13	103	103.5	103.0	1.1	0.20	234	234	233.8	0.4
0.16	137	138	137.0	2.2	0.24	191	194	192.5	2.1
0.19	132	132	131.5	1.1	0.29	122	125	123.5	2.1
0.21	127	127	127.0	0.0	0.33	112	113	112.5	0.7
0.24	128	128	128.0	0.0	0.38	115	116	115.5	0.7
0.27	124	124	124.0	0.0	0.42	110	112	111.0	1.4
0.30	123	124	123.0	2.2	0.46	102	103	102.5	0.7
0.33	137	136.5	136.5	0.0	0.51	104	105	104.5	0.7
0.36	148	148	148.0	0.0	0.55	121	123	122.0	1.4
0.39	149	149.5	149.0	1.1	0.60	134	135	134.5	0.7
0.41	149	149	149.0	0.0	0.64	131	132	131.5	0.7
0.44	156	156	156.0	0.0	0.69	95.5	95	95.3	0.4
0.47	123	124	123.0	2.2	0.73	62	64	63.0	1.4
0.50	84	84	84.0	0.0	0.77	50.5	52	51.3	1.1
0.53	38	38	38.0	0.0	0.82	33	34	33.5	0.7
					0.86	25.5	27	26.3	1.1
					0.91	27	27	27.0	0.0
					0.95	30	32	31.0	1.4
					1.00	27	28	27.5	0.7

1.04	25	26	25.5	0.7
1.08	21	22	21.5	0.7
1.13	23	24	23.5	0.7
1.17	23.5	25	24.3	1.1
1.22	22	23	22.5	0.7
1.26	20	21	20.5	0.7
1.31	19	20.5	19.8	1.1
1.35	14	16	15.0	1.4
1.39	10	11	10.5	0.7
1.44	9	10	9.5	0.7
1.48	10	11	10.5	0.7
1.53	11	12	11.5	0.7
1.57	11	12	11.5	0.7
1.61	11	12	11.5	0.7
1.66	11	13	12.0	1.4
1.70	12	13	12.5	0.7
1.75	12	13	12.5	0.7
1.79	11	13	12.0	1.4
1.84	11	12	11.5	0.7
1.88	9	10	9.5	0.7

Core TCB4

Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.02	36	36	36.0	0.0
0.07	45	45	45.0	0.0
0.12	50	50	50.0	0.0
0.16	60	60	60.0	0.0
0.21	96	97	96.5	2.2
0.26	141	141	140.8	1.1
0.30	139	139	139.0	0.0
0.35	152	153	152.5	2.2
0.40	178	178	178.0	0.0
0.44	171	169	170.0	4.3
0.49	147	146	146.5	2.2
0.54	139	138	138.3	1.1
0.58	116	117	116.3	3.2
0.63	77	77	77.0	0.0
0.68	59	59	59.0	0.0
0.72	50	50	50.0	0.0
0.77	50	50	50.0	0.0
0.82	62	61	61.5	2.2
0.86	71	71	71.0	0.0
0.91	67	67	67.0	0.0
0.96	56.5	57	56.8	1.1
1.00	44	44	44.0	0.0
1.05	30	30	30.0	0.0
1.10	25	24	24.5	2.2
1.15	21	21	21.0	0.0
1.19	18	17	17.5	2.2
1.24	15	14	14.5	2.2
1.29	12	13	12.5	2.2
1.33	12	11	11.5	2.2
1.38	12	12	12.0	0.0
1.43	12	12	12.0	0.0
1.47	12.5	12	12.3	1.1
1.52	13	13	13.0	0.0
1.57	13	13	13.0	0.0
1.61	13	13	13.0	0.0
1.66	14	14	14.0	0.0

Core TCB5

Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.03	28	31	29.5	6.5
0.10	44	46	45.0	4.3
0.17	60	62	61.0	4.3
0.23	78.5	80	79.3	3.2
0.30	99	100	99.5	2.2
0.37	107	107	107.0	0.0
0.43	102	101	101.3	1.1
0.50	99	99	99.0	0.0
0.57	84	85	84.5	2.2
0.63	64	64	64.0	0.0
0.70	61	61	61.0	0.0
0.77	60	60	60.0	0.0
0.83	58	58	58.0	0.0
0.90	51	51	51.0	0.0
0.97	40	39	39.5	2.2
1.03	28	29	28.5	2.2
1.10	24	24	24.0	0.0
1.17	21	21	21.0	0.0
1.23	19	19	19.0	0.0
1.30	18	18	18.0	0.0
1.37	17	17	17.0	0.0

Core TCB6

Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.03	82	81	81.5	2.2
0.08	120	120	120.0	0.0
0.13	84	84	84.0	0.0
0.18	50	50	50.0	0.0
0.23	41	41	41.0	0.0
0.28	38	38	38.0	0.0
0.33	36	36	36.0	0.0
0.38	35	35	35.0	0.0
0.43	34	34	34.0	0.0

Core TCB7

Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.03	29	28	28.5	2.2
0.09	37	36	36.5	2.2
0.15	43.5	43	43.3	1.1
0.21	47.5	48	47.8	1.1
0.27	59	58	58.5	2.2
0.34	81	81	81.0	0.0
0.40	95	95	95.0	0.0
0.46	83	83	83.0	0.0
0.52	72	71	71.5	2.2

0.48	31	31	31.0	0.0	0.58	73	73	73.0	0.0
0.53	30	29.5	29.8	1.1	0.64	77	77	77.0	0.0
0.58	31	31	31.0	0.0	0.70	86	86	86.0	0.0
0.63	33	33	33.0	0.0	0.76	90	90	90.0	0.0
0.68	30	30	30.0	0.0	0.82	63	61	62.0	4.3
0.73	29	29	29.0	0.0	0.89	51	51	51.0	0.0
0.78	29	29	29.0	0.0	0.95	59.5	59	59.3	1.1
0.83	28	28	28.0	0.0	1.01	59	60	59.5	2.2
					1.07	43	43	43.0	0.0
					1.13	43	42	42.5	2.2
					1.19	41	41	41.0	0.0
					1.25	41	41	41.0	0.0
					1.31	46	46	46.0	0.0
					1.37	53	53	53.0	0.0
					1.44	47	46	46.5	2.2
					1.50	35	35	35.0	0.0
					1.56	25	25	25.0	0.0
					1.62	21	21	21.0	0.0
					1.68	25	25	25.0	0.0
					1.74	24.5	24	24.3	1.1
					1.80	18	18	18.0	0.0
					1.86	13	13	13.0	0.0
					1.92	12	12	12.0	0.0
					1.99	13	13	13.0	0.0
					2.05	14	14	14.0	0.0
					2.11	16	15	15.5	2.2
					2.17	16	16	16.0	0.0
					2.23	15	16	15.5	2.2
					2.29	11	11	11.0	0.0

Core TCB8					**Core TCC8**				
Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error	Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.03	24	24	24.0	0.0	0.04	58	59	58.5	2.2
0.10	26	26	26.0	0.0	0.11	84	84	84.0	0.0
0.16	26	26	26.0	0.0	0.19	56	57	56.5	2.2
0.23	26	25	25.3	1.1	0.26	44	44	44.0	0.0
0.29	28	28	28.0	0.0	0.34	44	44	44.0	0.0
0.35	43	43	43.0	0.0	0.41	45	45	45.0	0.0
0.42	71	70	70.5	2.2	0.49	55	54	54.5	2.2
0.48	66	66	66.0	0.0	0.56	126	126	126.0	0.0
0.55	51	51	51.0	0.0	0.64	360	362	361.0	4.3
0.61	54	54	54.0	0.0	0.71	265	267	266.0	4.3
0.68	65	65	65.0	0.0	0.79	83	81	82.0	4.3
0.74	71	71	71.0	0.0	0.86	53	54	53.5	2.2
0.81	76	76	76.0	0.0	0.94	44	44	44.0	0.0
0.87	82	82	81.8	1.1	1.01	39	39	39.0	0.0
0.93	76	76	75.8	1.1	1.05	35	34	34.5	2.2
1.00	58	58	58.0	0.0	1.09	26	25	25.5	2.2
1.06	45	45	45.0	0.0					
1.13	40	40	40.0	0.0					
1.19	37	37	37.0	0.0					
1.26	32	32.5	32.3	1.1					
1.32	28	28	28.0	0.0					
1.39	24	24	24.0	0.0					
1.45	21	22	21.5	2.2					
1.52	17	18	17.5	2.2					
1.58	13	13	13.0	0.0					
1.64	10	11	10.5	2.2					
1.71	10	11	10.5	2.2					
1.77	10	10	10.0	0.0					
1.84	11	11	11.0	0.0					
1.90	11	11	11.0	0.0					
1.97	10	11	10.5	2.2					
2.03	10	11	10.5	2.2					
2.10	11	11	11.0	0.0					
2.16	11	11	11.0	0.0					
2.22	12	12	12.0	0.0					
2.29	13	13	13.0	0.0					
2.35	14	15	14.5	2.2					
2.42	16	16	16.0	0.0					
2.48	15	15	15.0	0.0					

Core TCD8					**Core TCD9**				
Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error	Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
0.02	21	20	20.5	2.2	0.02	34	35	34.5	2.2
0.07	50	50	50.0	0.0	0.07	68	67	67.5	2.2
0.11	58	56	57.0	4.3	0.11	95	95	95.0	0.0
0.16	77	72	74.5	10.8	0.16	123	123	123.0	0.0
0.21	44	44	44.0	0.0	0.21	96	97	96.5	2.2
0.25	38	38	38.0	0.0	0.25	83	83.5	83.3	1.1
0.30	36	36	36.0	0.0	0.30	86	86	86.0	0.0
0.34	39	39	39.0	0.0	0.34	119	110	114.5	19.4
0.39	55	54	54.5	2.2	0.39	168	169	168.5	2.2
0.44	76	77	76.5	2.2	0.44	145	146	145.5	2.2
0.48	71	71	71.0	0.0	0.48	128	128	128.0	0.0
0.53	56	56	56.0	0.0	0.53	140	140	140.0	0.0
0.57	56	57	56.5	2.2	0.57	151	150	150.5	2.2
0.62	68	69	68.5	2.2	0.62	159	157	158.0	4.3
0.67	71	72	71.5	2.2	0.67	125	126	125.5	2.2
0.71	57	57	57.0	0.0	0.71	84	86	85.0	4.3
0.76	39	39	39.0	0.0	0.76	58	60	59.0	4.3
0.80	27	27	27.0	0.0	0.80	50	51	50.5	2.2
0.85	23	23	23.0	0.0	0.85	48	49	48.5	2.2
0.90	22	22	22.0	0.0	0.90	37	38	37.5	2.2
0.94	17	17	17.0	0.0	0.94	33	33	33.0	0.0
0.99	15	15	15.0	0.0					
1.03	15	15	15.0	0.0	**Core TCE9**				
1.08	14	14	14.0	0.0	Depth (m)	VSMS 1	VSMS 2	Mean VSMS	VSMS error
1.12	13	13.5	13.3	1.1	0.02	37	36	36.5	2.2
1.17	12	12.5	12.3	1.1	0.06	40	39	39.5	2.2
1.22	10	10	10.0	0.0	0.10	43	43	43.0	0.0
1.26	11	11	11.0	0.0	0.14	53	53	53.0	0.0
1.31	9	9.5	9.3	1.1	0.18	57	57	57.0	0.0
1.35	9	9.5	9.3	1.1	0.22	56	57	56.5	2.2
1.40	9	9	9.0	0.0	0.26	52.5	52	52.3	1.1
1.45	9	9.5	9.3	1.1	0.29	46	46	46.0	0.0
1.49	10	9.5	9.8	1.1	0.33	47	46	46.5	2.2
1.54	11	10.5	10.8	1.1	0.37	39	39	39.0	0.0
1.58	13	12	12.5	2.2	0.41	28	26	27.0	4.3
1.63	14	13.5	13.8	1.1	0.45	24	24	24.0	0.0
1.68	14	14	14.0	0.0	0.49	24	24	24.0	0.0
					0.53	23	22	22.5	2.2
					0.57	21	20	20.5	2.2
					0.61	20	19	19.5	2.2
					0.65	20	19	19.5	2.2
					0.69	18	18	18.0	0.0
					0.73	10	10	10.0	0.0

Appendix 2

The moisture content, dry bulk density, loss-on-ignition and dry mineral bulk density of core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

Upper depth	Lower depth	Moisture content	Dry bulk density	Loss-on-ignition	Dry mineral bulk density
(m)	(m)	(%)	(kg m^{-3})	(LOI_{430}) (%)	(kg m^{-3})
0.00	0.02	65.52	140.02	10.50	140.02
0.02	0.04	65.16	237.86	10.39	237.86
0.04	0.06	64.95	320.71	9.26	320.71
0.06	0.08	65.07	350.29	10.68	350.29
0.08	0.10	67.05	206.04	12.22	206.04
0.10	0.12	66.94	227.56	9.48	227.56
0.12	0.14	63.57	294.50	9.76	294.50
0.14	0.16	57.32	377.75	5.94	377.75
0.16	0.18	54.23	414.19	7.55	414.19
0.18	0.20	55.58	426.60	8.47	426.60
0.20	0.22	57.42	372.07	8.11	372.07
0.22	0.24	55.86	314.90	7.79	314.90
0.24	0.26	56.18	463.84	7.54	463.84
0.26	0.28	56.37	440.99	6.81	440.99
0.28	0.30	58.19	357.81	9.89	357.81
0.30	0.32	62.10	361.05	9.57	361.05
0.32	0.34	59.45	391.09	6.54	391.09
0.34	0.36	55.07	412.14	8.91	412.14
0.36	0.38	64.87	299.26	8.93	299.26
0.38	0.40	58.45	395.44	8.34	395.44
0.40	0.42	54.81	487.27	7.93	487.27
0.42	0.44	56.19	391.28	8.52	391.28
0.44	0.46	60.45	386.33	10.11	386.33
0.46	0.48	57.70	411.88	8.99	411.88
0.48	0.50	56.91	415.91	10.09	415.91
0.50	0.52	62.55	333.85	12.28	333.85
0.52	0.54	69.71	216.54	14.68	216.54
0.54	0.56	64.33	347.25	12.09	347.25
0.56	0.58	59.17	423.50	8.85	423.50
0.58	0.60	53.27	434.85	8.06	434.85
0.60	0.62	52.73	447.07	8.44	447.07
0.62	0.64	48.43	546.49	7.54	546.49
0.64	0.66	47.72	508.46	7.31	508.46
0.66	0.68	48.29	523.98	7.80	523.98
0.68	0.70	49.25	504.70	8.45	504.70
0.70	0.72	48.05	539.43	8.23	539.43
0.72	0.74	46.47	624.92	8.13	624.92
0.74	0.76	56.50	401.65	11.09	401.65
0.76	0.78	57.48	369.63	13.98	369.63
0.78	0.80	41.18	624.46	11.98	624.46
0.80	0.82	39.64	681.16	10.46	681.16
0.82	0.84	37.81	718.73	10.03	718.73
0.84	0.86	37.18	803.03	10.60	803.03
0.86	0.88	36.44	734.90	5.96	734.90
0.88	0.90	37.69	599.70	9.96	599.70
0.90	0.92	38.88	769.16	9.07	769.16
0.92	0.94	36.94	779.53	9.79	779.53

0.94	0.96	43.12	622.87	12.43	622.87
0.96	0.98	47.13	494.47	22.06	494.47
0.98	1.00	51.22	482.12	29.21	482.12
1.00	1.02	51.21	532.49	27.47	532.49
1.02	1.04	51.87	389.76	19.14	389.76
1.04	1.06	52.51	473.94	18.51	473.94
1.06	1.08	52.75	421.65	20.76	421.65
1.08	1.10	52.31	435.84	20.06	435.84
1.10	1.12	51.03	578.18	26.18	578.18
1.12	1.14	51.25	385.34	18.06	385.34
1.14	1.16	52.41	454.59	19.45	454.59
1.16	1.18	52.20	454.53	18.21	454.53
1.18	1.20	52.82	508.20	16.23	508.20
1.20	1.22	49.25	532.96	13.32	532.96
1.22	1.24	46.23	565.50	11.71	565.50
1.24	1.26	42.94	626.44	9.80	626.44
1.26	1.28	44.33	672.19	9.50	672.19
1.28	1.30	44.27	536.19	10.83	536.19
1.30	1.32	43.35	734.04	10.25	734.04
1.32	1.34	42.98	628.94	10.51	628.94
1.34	1.36	42.11	482.65	9.74	482.65
1.36	1.38	39.62	690.41	7.32	690.41
1.38	1.40	39.29	743.29	7.93	743.29
1.40	1.42	40.90	622.54	8.82	622.54
1.42	1.44	40.65	704.73	7.86	704.73
1.44	1.46	43.77	580.16	7.77	580.16
1.46	1.48	44.37	616.01	7.24	616.01
1.48	1.50	43.43	522.00	7.53	522.00
1.50	1.52	41.02	787.45	8.02	787.45
1.52	1.54	41.14	659.58	6.39	659.58
1.54	1.56	41.31	696.81	7.39	696.81
1.56	1.58	41.28	617.79	7.60	617.79
1.58	1.60	41.95	628.15	6.74	628.15
1.60	1.62	42.35	675.75	8.41	675.75
1.62	1.64	42.40	683.28	7.12	683.28
1.64	1.66	40.86	620.03	8.82	620.03
1.66	1.68	37.98	683.14	7.67	683.14
1.68	1.70	36.74	684.73	7.67	684.73
1.70	1.72	37.07	676.01	7.86	676.01
1.72	1.74	37.12	847.86	6.92	847.86
1.74	1.76	37.66	678.79	8.09	678.79
1.76	1.78	39.00	678.52	9.00	678.52
1.78	1.80	38.96	758.47	9.13	758.47
1.80	1.82	40.10	741.97	9.79	741.97
1.82	1.84	41.27	566.62	8.72	566.62
1.84	1.86	41.88	670.93	9.72	670.93
1.86	1.88	40.31	621.75	8.94	621.75
1.88	1.90	38.87	693.25	9.09	693.25
1.90	1.92	38.02	745.33	9.02	745.33
1.92	1.94	37.17	647.89	8.82	647.89
1.94	1.96	36.76	864.16	8.41	864.16
1.96	1.98	37.07	644.00	8.16	644.00
1.98	2.00	37.29	795.70	6.75	795.70

Appendix 3

The mean low-frequency magnetic susceptibility (K_{lf}), mean high-frequency susceptibility (K_{hf}) and frequency-dependant magnetic susceptibility (χ_{fd}) of core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Measures of error represent 95% confidence limits of the measurements.

Upper depth	Lower depth	K_{lf}	K_{hf}	χ_{fd}	χ_{fd} error	Mean χ_{lf}	χ_{lf} error
(m)	(m)	(10^{-5} SI units)	(10^{-5} SI units)	(%)	(%)	($m^3\ kg^{-1}$)	($m^3\ kg^{-1}$)
0.00	0.02	179.0	178.0	0.56	0.00	172.4	0.0
0.02	0.04	161.3	159.0	1.45	0.71	180.9	1.3
0.04	0.06	241.5	240.0	0.62	0.66	223.9	1.5
0.06	0.08	356.3	354.0	0.87	0.46	379.2	1.5
0.08	0.10	139.1	136.0	1.83	1.07	129.4	0.9
0.10	0.12	78.7	78.0	0.85	1.82	78.3	1.4
0.12	0.14	128.9	124.0	3.19	0.98	113.4	0.9
0.14	0.16	144.3	140.0	2.69	0.93	151.5	1.2
0.16	0.18	150.8	146.0	3.37	0.97	163.5	1.3
0.18	0.20	113.3	109.0	3.16	1.19	119.0	1.1
0.20	0.22	88.5	86.0	2.37	1.25	92.4	0.9
0.22	0.24	52.8	51.0	3.79	1.88	65.5	0.7
0.24	0.26	73.3	71.0	3.46	1.62	64.4	0.8
0.26	0.28	72.0	70.0	2.43	1.85	70.4	1.0
0.28	0.30	59.4	59.0	1.94	1.76	69.2	0.8
0.30	0.32	39.3	39.0	1.78	2.73	34.9	0.7
0.32	0.34	59.0	58.0	1.69	0.00	48.7	0.0
0.34	0.36	68.6	67.0	2.00	1.74	64.2	0.8
0.36	0.38	61.4	60.0	2.28	1.81	52.6	1.0
0.38	0.40	101.2	99.0	2.14	0.78	80.6	0.6
0.40	0.42	129.0	126.0	2.33	0.89	87.6	0.8
0.42	0.44	70.3	70.0	0.97	1.78	58.9	0.7
0.44	0.46	58.6	57.0	2.73	1.89	49.6	0.9
0.46	0.48	55.0	54.0	3.18	1.99	46.3	0.6
0.48	0.50	53.1	52.0	1.68	2.17	42.6	0.7
0.50	0.52	39.2	38.0	2.13	2.45	36.2	0.7
0.52	0.54	32.5	33.0	0.00	0.00	25.9	1.3
0.54	0.56	34.0	33.0	0.00	0.00	33.0	0.9
0.56	0.58	60.0	57.0	2.78	1.43	50.9	0.0
0.58	0.60	72.0	70.0	1.74	1.10	61.5	0.0
0.60	0.62	58.3	57.0	2.00	2.00	48.5	0.7
0.62	0.64	56.6	54.0	3.58	1.76	36.5	0.4
0.64	0.66	16.5	15.0	5.05	10.95	21.0	2.0
0.66	0.68	18.1	18.0	4.83	6.80	22.6	1.2
0.68	0.70	20.8	21.0	2.08	6.73	25.5	1.5
0.70	0.72	25.8	25.0	4.19	5.66	29.4	1.4
0.72	0.74	23.0	21.0	6.09	4.83	23.4	0.9
0.74	0.76	31.4	31.0	0.00	0.00	25.7	0.9
0.76	0.78	32.0	31.0	3.13	2.74	33.4	0.9
0.78	0.80	33.4	31.0	2.99	3.90	36.2	0.7
0.80	0.82	33.4	32.0	1.78	4.24	34.6	1.2
0.82	0.84	38.3	37.0	1.31	2.94	35.7	0.7
0.84	0.86	36.4	36.0	0.00	0.00	32.5	0.6
0.86	0.88	24.4	24.0	2.78	4.93	33.9	1.3
0.88	0.90	21.3	21.0	0.00	0.00	31.2	1.6

0.90	0.92	21.2	20.0	2.68	5.84	21.0	1.0
0.92	0.94	15.0	13.0	5.00	4.94	16.2	0.0
0.94	0.96	12.6	12.0	0.00	0.00	16.1	1.4
0.96	0.98	9.3	9.0	0.00	0.00	15.0	2.3
0.98	1.00	15.5	16.0	0.00	0.00	14.5	1.5
1.00	1.02	9.4	10.0	0.00	0.00	15.1	1.4
1.02	1.04	14.0	15.0	0.00	0.00	14.7	0.9
1.04	1.06	9.3	8.0	5.41	12.16	15.5	1.3
1.06	1.08	12.8	13.0	0.00	0.00	12.6	0.8
1.08	1.10	12.4	12.0	0.00	0.00	13.5	0.7
1.10	1.12	11.0	11.0	0.00	0.00	13.6	1.1
1.12	1.14	9.5	10.0	0.00	0.00	12.4	2.1
1.14	1.16	6.6	6.0	9.09	16.82	12.2	2.1
1.16	1.18	6.6	8.0	0.00	0.00	10.9	1.8
1.18	1.20	8.6	8.0	5.81	10.01	12.7	1.0
1.20	1.22	8.6	8.0	6.98	12.91	12.5	1.6
1.22	1.24	9.1	9.0	0.00	0.00	11.9	1.1
1.24	1.26	12.6	12.0	0.00	0.00	13.8	0.7
1.26	1.28	8.4	7.0	5.08	16.23	10.0	1.2
1.28	1.30	5.0	5.0	10.00	31.82	7.3	0.0
1.30	1.32	7.4	8.0	0.00	0.00	7.5	0.7
1.32	1.34	6.7	6.0	10.00	12.86	7.8	1.0
1.34	1.36	5.7	4.0	18.82	15.35	8.2	0.8
1.36	1.38	7.0	6.0	11.43	17.20	7.7	1.2
1.38	1.40	5.5	6.0	0.00	0.00	6.7	1.9
1.40	1.42	4.7	5.0	0.00	0.00	5.9	1.1
1.42	1.44	5.8	5.0	17.24	25.39	7.6	1.8
1.44	1.46	8.6	8.0	9.88	10.48	7.0	0.6
1.46	1.48	8.0	7.0	0.00	0.00	6.6	0.0
1.48	1.50	7.3	8.0	0.00	0.00	6.0	0.9
1.50	1.52	7.6	6.0	9.36	11.94	7.3	0.7
1.52	1.54	8.0	8.0	0.00	0.00	6.7	0.7
1.54	1.56	4.8	5.0	0.00	0.00	6.3	1.0
1.56	1.58	7.0	7.0	0.00	0.00	7.0	0.9
1.58	1.60	8.3	9.0	0.00	0.00	6.6	0.6
1.60	1.62	7.0	8.0	0.00	0.00	6.0	0.0
1.62	1.64	7.2	8.0	0.00	0.00	6.4	0.7
1.64	1.66	7.0	7.0	0.00	0.00	6.4	0.7
1.66	1.68	5.3	5.0	0.00	0.00	6.9	1.0
1.68	1.70	5.3	4.0	23.81	15.15	7.2	1.1
1.70	1.72	5.6	6.0	0.00	0.00	6.9	0.8
1.72	1.74	5.8	6.0	0.00	0.00	6.5	0.9
1.74	1.76	5.4	6.0	0.00	0.00	6.6	0.9
1.76	1.78	5.8	5.0	0.00	0.00	7.5	1.3
1.78	1.80	5.5	5.0	9.09	16.25	7.3	1.2
1.80	1.82	5.7	6.0	0.00	0.00	6.8	0.7
1.82	1.84	7.5	7.6	0.00	0.00	7.3	0.7
1.84	1.86	4.5	4.9	0.00	0.00	6.9	1.0
1.86	1.88	8.1	8.4	0.00	0.00	7.3	0.3
1.88	1.90	6.1	6.1	0.00	0.00	7.8	0.6
1.90	1.92	6.6	6.9	0.00	0.00	7.5	0.4
1.92	1.94	6.4	6.7	0.00	0.00	7.5	0.6
1.94	1.96	8.6	8.4	0.00	0.00	7.9	0.5
1.96	1.98	9.0	8.8	0.00	0.00	7.9	0.8
1.98	2.00	5.7	5.9	0.00	0.00	7.4	0.9

Appendix 4

The mean natural remanent magnetisation (NRM), mean anhysteretic remanent magnetisation (ARM) and mean mass-specific anhysteretic remanent magnetisation (χ_{ARM}) of core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Errors represent 95% confidence limits of the measurements.

Upper depth	Lower depth	Mean NRM	NRM error	Mean ARM	ARM error	Mean χ_{ARM}	χ_{ARM} error
(m)	(m)	(10^{-6} Am^2kg^{-1})	(10^{-6} Am^2kg^{-1})	(10^{-5} Am^2kg^{-1})	(10^{-5} Am^2kg^{-1})	(10^{-7} Am^2kg^{-1})	(10^{-7} Am^2kg^{-1})
0.00	0.02	55.99	5.47	13.23	0.53	41.5	1.66
0.02	0.04	51.06	7.67	13.28	0.75	41.7	2.36
0.04	0.06	23.28	3.04	13.55	0.44	42.6	1.38
0.06	0.08	66.16	10.36	15.42	0.23	48.4	0.72
0.08	0.10	59.31	2.78	15.98	0.18	50.2	0.58
0.10	0.12	16.66	0.73	16.07	0.24	50.5	0.76
0.12	0.14	14.62	0.62	16.06	0.74	50.5	2.31
0.14	0.16	10.27	0.76	18.39	0.04	57.8	0.12
0.16	0.18	17.00	0.18	18.12	0.15	56.9	0.47
0.18	0.20	6.13	0.38	14.81	0.03	46.5	0.10
0.20	0.22	9.62	0.40	13.10	0.01	41.2	0.02
0.22	0.24	5.75	0.16	9.33	0.04	29.3	0.12
0.24	0.26	13.76	0.20	9.13	0.09	28.7	0.27
0.26	0.28	14.29	0.09	9.64	0.02	30.3	0.06
0.28	0.30	24.19	0.97	7.42	0.24	23.3	0.77
0.30	0.32	5.12	0.00	4.46	0.04	14.0	0.12
0.32	0.34	9.02	0.06	6.30	0.22	19.8	0.68
0.34	0.36	6.43	0.14	6.46	0.15	20.3	0.48
0.36	0.38	7.22	0.05	7.03	0.07	22.1	0.22
0.38	0.40	10.49	0.02	10.79	0.08	33.9	0.25
0.40	0.42	12.95	0.20	13.05	0.00	41.0	0.00
0.42	0.44	8.95	0.04	10.20	0.16	32.0	0.52
0.44	0.46	6.03	0.02	12.69	0.27	39.9	0.85
0.46	0.48	8.41	0.08	10.78	0.06	33.9	0.20
0.48	0.50	7.78	0.12	8.23	0.19	25.8	0.59
0.50	0.52	5.91	0.04	6.84	0.06	21.5	0.18
0.52	0.54	4.54	0.15	4.68	0.04	14.7	0.14
0.54	0.56	3.25	0.13	4.28	0.06	13.4	0.18
0.56	0.58	8.80	0.02	9.53	0.01	29.9	0.03
0.58	0.60	12.99	0.10	10.07	0.17	31.6	0.55
0.60	0.62	10.60	0.24	9.17	0.21	28.8	0.67
0.62	0.64	4.81	0.06	6.59	0.12	20.7	0.36
0.64	0.66	2.97	0.14	4.73	0.08	14.8	0.25
0.66	0.68	2.73	0.03	6.04	0.07	19.0	0.23
0.68	0.70	3.83	0.05	6.42	0.15	20.2	0.47
0.70	0.72	5.22	0.07	6.67	0.00	21.0	0.01
0.72	0.74	2.30	0.05	6.14	0.10	19.3	0.31
0.74	0.76	4.75	0.10	9.48	0.01	29.8	0.03
0.76	0.78	7.04	0.13	8.82	0.04	27.7	0.11
0.78	0.80	5.08	0.00	4.09	0.16	12.9	0.49
0.80	0.82	32.76	0.25	4.27	0.02	13.4	0.07
0.82	0.84	9.02	0.04	3.91	0.09	12.3	0.29
0.84	0.86	2.63	0.09	3.39	0.06	10.6	0.17
0.86	0.88	3.82	0.00	3.29	0.01	10.3	0.04
0.88	0.90	2.37	0.00	3.08	0.02	9.7	0.08

0.90	0.92	1.77	0.00	2.01	0.01	6.3	0.03
0.92	0.94	1.66	0.22	2.26	0.05	7.1	0.15
0.94	0.96	1.39	0.08	1.79	0.01	5.6	0.03
0.96	0.98	1.32	0.10	1.88	0.16	5.9	0.50
0.98	1.00	2.12	0.11	1.75	0.09	5.5	0.27
1.00	1.02	2.00	0.12	1.76	0.03	5.5	0.10
1.02	1.04	1.44	0.13	1.66	0.14	5.2	0.44
1.04	1.06	1.89	0.20	1.37	0.00	4.3	0.00
1.06	1.08	2.01	0.10	1.57	0.01	4.9	0.04
1.08	1.10	2.59	0.02	1.72	0.04	5.4	0.13
1.10	1.12	4.03	0.05	1.71	0.04	5.4	0.13
1.12	1.14	2.30	0.19	1.48	0.04	4.7	0.13
1.14	1.16	1.93	0.08	1.37	0.00	4.3	0.00
1.16	1.18	1.93	0.20	1.40	0.00	4.4	0.00
1.18	1.20	1.40	0.05	1.35	0.00	4.2	0.00
1.20	1.22	1.05	0.11	1.23	0.00	3.9	0.00
1.22	1.24	1.10	0.09	1.18	0.00	3.7	0.00
1.24	1.26	1.30	0.04	1.34	0.00	4.2	0.00
1.26	1.28	0.48	0.02	0.97	0.01	3.0	0.02
1.28	1.30	0.21	0.08	0.68	0.01	2.1	0.02
1.30	1.32	0.34	0.10	0.68	0.06	2.1	0.18
1.32	1.34	0.29	0.09	0.68	0.02	2.1	0.05
1.34	1.36	0.49	0.17	0.60	0.06	1.9	0.18
1.36	1.38	0.38	0.16	0.60	0.01	1.9	0.03
1.38	1.40	0.13	0.08	0.59	0.01	1.9	0.03
1.40	1.42	0.21	0.04	0.57	0.00	1.8	0.00
1.42	1.44	0.21	0.04	0.56	0.00	1.7	0.00
1.44	1.46	0.24	0.00	0.50	0.00	1.6	0.00
1.46	1.48	0.08	0.04	0.53	0.00	1.7	0.00
1.48	1.50	0.05	0.12	0.52	0.00	1.6	0.00
1.50	1.52	0.47	0.04	0.54	0.00	1.7	0.00
1.52	1.54	0.75	0.30	0.55	0.00	1.7	0.00
1.54	1.56	0.52	0.04	0.52	0.00	1.6	0.00
1.56	1.58	0.25	0.00	0.55	0.00	1.7	0.00
1.58	1.60	0.60	0.00	0.52	0.00	1.6	0.00
1.60	1.62	2.20	0.31	0.46	0.00	1.4	0.00
1.62	1.64	0.73	0.00	0.41	0.00	1.3	0.00
1.64	1.66	0.55	0.00	0.40	0.02	1.2	0.06
1.66	1.68	0.41	0.00	0.43	0.00	1.4	0.00
1.68	1.70	0.46	0.16	0.45	0.00	1.4	0.00
1.70	1.72	0.44	0.00	0.44	0.00	1.4	0.00
1.72	1.74	0.45	0.00	0.46	0.00	1.4	0.00
1.74	1.76	0.50	0.00	0.45	0.00	1.4	0.00
1.76	1.78	0.50	0.00	0.43	0.00	1.3	0.00
1.78	1.80	0.49	0.00	0.43	0.00	1.3	0.00
1.80	1.82	0.66	0.00	0.44	0.00	1.4	0.00
1.82	1.84	0.57	0.00	0.42	0.00	1.3	0.00
1.84	1.86	0.39	0.22	0.41	0.00	1.3	0.00
1.86	1.88	0.38	0.00	0.42	0.00	1.3	0.00
1.88	1.90	0.48	0.00	0.43	0.00	1.4	0.00
1.90	1.92	0.44	0.00	0.42	0.00	1.3	0.00
1.92	1.94	0.45	0.17	0.41	0.00	1.3	0.00
1.94	1.96	0.41	0.00	0.44	0.00	1.4	0.00
1.96	1.98	0.50	0.04	0.45	0.00	1.4	0.00
1.98	2.00	0.43	0.00	0.52	0.00	1.6	0.00

Appendix 5

The mean isothermal remanent magnetisation ($IRM_{100\,mT}$) and mean saturated isothermal remanent magnetisation (SIRM = $IRM_{1000\,mT}$) of core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Errors represent 95% confidence limits of the measurements.

Upper depth	Lower depth	Mean $IRM_{100\,mT}$	$IRM_{100\,mT}$ error	Mean SIRM	SIRM Error
(m)	(m)	(10^{-5} A m^2kg^{-1})	(10^{-5} A m^2kg^{-1})	(10^{-5} A m^2kg^{-1})	(10^{-5} A m^2kg^{-1})
0.00	0.02	1625.1	44.7	1818.6	43.4
0.02	0.04	1583.8	48.6	1889.5	84.1
0.04	0.06	1542.6	83.0	1957.0	146.9
0.06	0.08	2122.6	144.5	2495.9	44.6
0.08	0.10	1255.0	126.4	1492.9	90.3
0.10	0.12	1172.1	75.6	1400.7	12.7
0.12	0.14	1510.8	208.9	1801.6	127.0
0.14	0.16	2094.6	96.9	2568.8	90.1
0.16	0.18	2264.9	138.0	2717.8	161.6
0.18	0.20	1606.7	62.5	1961.3	59.9
0.20	0.22	1519.3	37.6	1772.6	162.3
0.22	0.24	883.0	16.8	1123.5	7.2
0.24	0.26	873.5	46.4	1109.6	35.4
0.26	0.28	997.4	73.5	1254.9	84.4
0.28	0.30	921.5	85.9	1114.5	203.8
0.30	0.32	453.9	8.2	544.5	74.0
0.32	0.34	681.9	80.8	819.5	125.6
0.34	0.36	890.7	137.3	1038.1	26.5
0.36	0.38	807.6	135.0	968.4	88.1
0.38	0.40	1325.6	117.9	1586.9	88.2
0.40	0.42	1466.4	99.6	1773.5	22.3
0.42	0.44	912.4	101.9	1078.7	39.5
0.44	0.46	759.0	106.3	878.5	14.2
0.46	0.48	684.4	20.6	815.2	20.1
0.48	0.50	628.0	47.5	757.8	28.9
0.50	0.52	550.6	36.2	689.8	19.4
0.52	0.54	381.6	56.4	483.2	3.7
0.54	0.56	446.2	29.3	534.4	20.7
0.56	0.58	780.4	23.6	924.2	10.5
0.58	0.60	925.7	81.5	1080.7	38.0
0.60	0.62	792.0	14.5	931.6	6.1
0.62	0.64	563.3	49.2	668.3	1.5
0.64	0.66	234.8	33.2	294.4	8.7
0.66	0.68	274.1	21.8	341.4	2.6
0.68	0.70	338.7	38.3	414.1	7.7
0.70	0.72	384.9	38.0	454.3	15.2
0.72	0.74	299.8	26.7	362.0	2.3
0.74	0.76	414.6	32.6	469.2	11.5
0.76	0.78	497.2	37.8	570.4	22.6
0.78	0.80	477.9	26.7	553.5	12.7
0.80	0.82	470.7	32.9	550.9	17.7
0.82	0.84	483.0	4.2	565.4	1.7
0.84	0.86	439.2	19.6	512.0	27.8
0.86	0.88	445.5	15.7	520.9	2.2
0.88	0.90	405.4	0.9	470.3	0.5
0.90	0.92	219.5	32.5	261.5	12.0

0.92	0.94	210.2	13.8	260.7	14.0
0.94	0.96	153.0	6.9	186.9	12.7
0.96	0.98	134.2	2.2	163.2	5.5
0.98	1.00	126.1	9.3	153.2	4.4
1.00	1.02	128.7	10.9	154.7	18.4
1.02	1.04	124.2	4.7	151.4	0.4
1.04	1.06	106.7	8.6	129.9	10.5
1.06	1.08	103.1	2.7	124.6	0.9
1.08	1.10	106.0	0.3	127.2	6.7
1.10	1.12	103.1	3.4	122.5	8.6
1.12	1.14	101.9	7.0	123.7	12.7
1.14	1.16	87.2	0.8	106.7	12.8
1.16	1.18	86.8	5.5	109.1	7.4
1.18	1.20	88.7	2.9	113.4	2.5
1.20	1.22	80.6	1.1	110.2	0.6
1.22	1.24	83.7	4.8	117.2	7.6
1.24	1.26	100.2	4.6	141.1	1.8
1.26	1.28	51.0	4.3	76.5	1.5
1.28	1.30	27.2	0.0	45.1	1.0
1.30	1.32	23.9	1.1	41.1	0.6
1.32	1.34	22.8	2.1	39.5	0.7
1.34	1.36	23.8	10.0	41.8	3.4
1.36	1.38	23.2	0.1	39.4	2.8
1.38	1.40	23.8	0.7	39.2	0.7
1.40	1.42	25.2	3.2	39.8	1.3
1.42	1.44	23.0	5.6	36.7	1.2
1.44	1.46	20.9	3.3	30.8	1.7
1.46	1.48	20.8	1.4	30.1	2.6
1.48	1.50	21.9	2.8	32.1	2.4
1.50	1.52	22.3	0.3	32.4	2.9
1.52	1.54	22.2	2.9	33.4	1.7
1.54	1.56	21.1	1.3	32.1	0.2
1.56	1.58	24.6	1.9	34.6	2.9
1.58	1.60	23.7	4.7	32.7	3.4
1.60	1.62	22.3	1.8	31.1	0.9
1.62	1.64	19.0	0.4	29.5	0.9
1.64	1.66	19.7	0.2	27.9	1.3
1.66	1.68	20.3	2.6	29.7	0.0
1.68	1.70	20.8	0.5	29.9	1.8
1.70	1.72	19.9	3.2	30.1	1.3
1.72	1.74	18.1	2.7	30.6	3.3
1.74	1.76	19.4	2.9	31.2	3.5
1.76	1.78	19.0	2.9	29.4	2.8
1.78	1.80	19.2	0.0	29.0	1.1
1.80	1.82	18.5	4.2	29.7	3.8
1.82	1.84	19.0	2.9	27.5	1.3
1.84	1.86	19.1	1.6	32.4	1.1
1.86	1.88	17.9	1.4	29.1	2.9
1.88	1.90	18.2	1.4	28.2	1.8
1.90	1.92	18.2	1.4	28.4	0.0
1.92	1.94	18.6	1.2	26.8	1.9
1.94	1.96	18.2	1.2	26.9	1.2
1.96	1.98	18.6	1.4	29.8	2.1
1.98	2.00	18.6	1.6	34.1	1.2

Appendix 6

The χ_{ARM}/χ_{lf}, S-ratio ($IRM_{100\,mT}/SIRM$), high-field IRM ($SIRM-IRM_{100\,mT}$), $SIRM/\chi_{lf}$, $\chi_{ARM}/SIRM$ ratios from core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia.

Upper depth	Lower depth	χ_{ARM}/χ_{lf}	S-Ratio	High-field IRM	$SIRM/\chi_{lf}$	$\chi_{ARM}/SIRM$
(m)	(m)	(dimensionless)	(dimensionless)	(10^{-5} A m^2kg^{-1})	(A m^{-1})	(10^{-5} m A^{-1})
0.00	0.02	2.41	0.89	193.5	1.05	22.84
0.02	0.04	2.31	0.84	305.7	1.04	22.07
0.04	0.06	1.90	0.79	414.4	0.87	21.75
0.06	0.08	1.28	0.85	373.4	0.66	19.40
0.08	0.10	3.88	0.84	237.9	1.15	33.62
0.10	0.12	6.45	0.84	228.6	1.79	36.03
0.12	0.14	4.45	0.84	290.8	1.59	28.00
0.14	0.16	3.81	0.82	474.1	1.70	22.49
0.16	0.18	3.48	0.83	452.9	1.66	20.94
0.18	0.20	3.91	0.82	354.6	1.65	23.72
0.20	0.22	4.45	0.86	253.3	1.92	23.22
0.22	0.24	4.48	0.79	240.5	1.72	26.08
0.24	0.26	4.45	0.79	236.1	1.72	25.84
0.26	0.28	4.30	0.79	257.5	1.78	24.12
0.28	0.30	3.36	0.83	193.0	1.61	20.90
0.30	0.32	4.01	0.83	90.6	1.56	25.72
0.32	0.34	4.07	0.83	137.7	1.68	24.15
0.34	0.36	3.16	0.86	147.4	1.62	19.54
0.36	0.38	4.20	0.83	160.8	1.84	22.79
0.38	0.40	4.20	0.84	261.3	1.97	21.35
0.40	0.42	4.68	0.83	307.1	2.02	23.12
0.42	0.44	5.44	0.85	166.4	1.83	29.71
0.44	0.46	8.03	0.86	119.5	1.77	45.37
0.46	0.48	7.31	0.84	130.9	1.76	41.52
0.48	0.50	6.06	0.83	129.8	1.78	34.10
0.50	0.52	5.93	0.80	139.2	1.91	31.14
0.52	0.54	5.68	0.79	101.5	1.87	30.41
0.54	0.56	4.08	0.83	88.2	1.62	25.14
0.56	0.58	5.88	0.84	143.8	1.82	32.38
0.58	0.60	5.14	0.86	155.0	1.76	29.28
0.60	0.62	5.94	0.85	139.6	1.92	30.92
0.62	0.64	5.68	0.84	105.0	1.83	30.97
0.64	0.66	7.06	0.80	59.6	1.40	50.43
0.66	0.68	8.39	0.80	67.3	1.51	55.58
0.68	0.70	7.92	0.82	75.4	1.63	48.73
0.70	0.72	7.14	0.85	69.4	1.55	46.14
0.72	0.74	8.26	0.83	62.2	1.55	53.28
0.74	0.76	11.59	0.88	54.6	1.83	63.47
0.76	0.78	8.29	0.87	73.2	1.71	48.56
0.78	0.80	3.56	0.86	75.7	1.53	23.23
0.80	0.82	3.88	0.85	80.2	1.59	24.35
0.82	0.84	3.44	0.85	82.4	1.58	21.71
0.84	0.86	3.28	0.86	72.8	1.58	20.79
0.86	0.88	3.05	0.86	75.4	1.54	19.83
0.88	0.90	3.11	0.86	64.9	1.51	20.59
0.90	0.92	3.02	0.84	42.0	1.25	24.18

0.92	0.94	4.39	0.81	50.4	1.61	27.26
0.94	0.96	3.49	0.82	33.9	1.16	30.09
0.96	0.98	3.93	0.82	29.1	1.09	36.10
0.98	1.00	3.79	0.82	27.2	1.06	35.79
1.00	1.02	3.67	0.83	26.1	1.03	35.78
1.02	1.04	3.53	0.82	27.2	1.03	34.41
1.04	1.06	2.78	0.82	23.3	0.84	33.16
1.06	1.08	3.90	0.83	21.5	0.99	39.55
1.08	1.10	4.01	0.83	21.2	0.94	42.49
1.10	1.12	3.95	0.84	19.4	0.90	43.90
1.12	1.14	3.74	0.82	21.8	0.99	37.61
1.14	1.16	3.51	0.82	19.5	0.87	40.17
1.16	1.18	4.05	0.80	22.3	1.00	40.41
1.18	1.20	3.33	0.78	24.7	0.89	37.32
1.20	1.22	3.10	0.73	29.6	0.88	35.14
1.22	1.24	3.11	0.71	33.5	0.98	31.68
1.24	1.26	3.05	0.71	40.9	1.02	29.84
1.26	1.28	3.05	0.67	25.5	0.77	39.74
1.28	1.30	2.90	0.60	17.9	0.61	47.12
1.30	1.32	2.85	0.58	17.2	0.55	51.69
1.32	1.34	2.71	0.58	16.7	0.50	53.74
1.34	1.36	2.28	0.57	18.0	0.51	44.96
1.36	1.38	2.45	0.59	16.2	0.51	47.72
1.38	1.40	2.77	0.61	15.4	0.58	47.44
1.40	1.42	3.03	0.63	14.6	0.68	44.64
1.42	1.44	2.31	0.63	13.7	0.49	47.53
1.44	1.46	2.24	0.68	9.9	0.44	51.14
1.46	1.48	2.55	0.69	9.3	0.46	55.72
1.48	1.50	2.71	0.68	10.3	0.54	50.51
1.50	1.52	2.32	0.69	10.1	0.44	52.36
1.52	1.54	2.59	0.66	11.3	0.50	51.84
1.54	1.56	2.61	0.66	11.0	0.51	51.01
1.56	1.58	2.47	0.71	10.0	0.49	50.29
1.58	1.60	2.46	0.72	9.0	0.50	49.48
1.60	1.62	2.41	0.72	8.8	0.52	46.55
1.62	1.64	2.04	0.65	10.4	0.46	43.97
1.64	1.66	1.94	0.71	8.1	0.43	44.60
1.66	1.68	1.98	0.68	9.3	0.43	45.66
1.68	1.70	1.97	0.69	9.2	0.42	47.47
1.70	1.72	2.01	0.66	10.2	0.44	45.81
1.72	1.74	2.21	0.59	12.6	0.47	47.04
1.74	1.76	2.13	0.62	11.8	0.47	44.99
1.76	1.78	1.79	0.65	10.4	0.39	45.52
1.78	1.80	1.85	0.66	9.8	0.40	46.34
1.80	1.82	2.01	0.62	11.2	0.43	46.28
1.82	1.84	1.84	0.69	8.5	0.38	48.47
1.84	1.86	1.87	0.59	13.4	0.47	39.89
1.86	1.88	1.81	0.62	11.2	0.40	45.55
1.88	1.90	1.75	0.65	10.0	0.36	48.08
1.90	1.92	1.73	0.64	10.2	0.38	45.89
1.92	1.94	1.73	0.69	8.2	0.36	48.17
1.94	1.96	1.72	0.68	8.8	0.34	50.17
1.96	1.98	1.79	0.62	11.2	0.38	47.60
1.98	2.00	2.19	0.55	15.5	0.46	47.66

Appendix 7

The whole-core natural remanence and magnetic inclination measurements made on cores TCA9c and TCA9d, Tocal Homestead Lagoon, Paterson, eastern Australia.

Core TCA9c				
Depth	**X-coordinate**	**Y-coordinate**	**Z-coordinate**	**Inclination**
(m)				(°)
0.00	11.5518	10.68604	-82.5176	-75.8
0.04	15.53682	13.5051	-92.9496	-73.6
0.09	20.13422	16.0994	-103.223	-71.7
0.13	23.91952	18.27611	-111.598	-70.4
0.18	27.63207	20.25652	-120.178	-69.4
0.22	29.5578	22.01917	-126.055	-68.9
0.26	32.52185	24.24489	-134.894	-68.3
0.31	34.01196	26.07512	-139.733	-67.9
0.35	35.16457	27.85938	-142.901	-67.4
0.40	36.03089	28.27489	-143.603	-67.1
0.44	35.60814	29.28671	-143.531	-66.9
0.48	35.1996	29.51706	-144.053	-67.1
0.53	34.60334	30.41875	-146.216	-67.3
0.57	34.78017	31.39219	-150.66	-67.6
0.62	36.07928	32.16961	-157.053	-67.8
0.66	37.54498	33.2984	-163.171	-67.8
0.70	39.78579	34.10508	-169.328	-67.7
0.75	41.46145	34.7795	-173.047	-67.5
0.79	43.35207	35.79065	-176.575	-67.1
0.84	44.29584	36.42117	-178.348	-66.9
0.88	44.75465	38.36156	-181.005	-66.6
0.92	45.42864	38.86298	-182.451	-66.5
0.97	44.75841	41.50126	-183.747	-66.2
1.01	44.01766	43.17348	-183.609	-65.9
1.06	41.79024	46.44632	-182.286	-65.4
1.10	39.12256	49.56703	-180.912	-65.0
1.14	35.2999	53.43647	-179.405	-64.4
1.19	32.52626	56.24171	-178.499	-63.9
1.23	28.84208	59.68519	-178.155	-63.4
1.28	25.94108	61.9627	-178.636	-63.1
1.32	21.85211	64.08424	-180.047	-63.0
1.36	20.16863	64.50337	-180.694	-63.1
1.41	18.03952	65.07539	-180.009	-63.0
1.45	17.86108	65.26775	-177.664	-62.7
1.49	17.03899	66.08782	-173.93	-62.0
1.54	17.05852	66.50282	-170.868	-61.4
1.58	18.62729	65.9425	-167.331	-61.0
1.63	19.06252	64.91031	-163.289	-60.7
1.67	19.45362	62.5273	-155.077	-60.2
1.71	19.27492	59.76476	-145.77	-59.7
1.76	18.35315	54.547	-131.111	-59.3
1.80	16.73749	50.2719	-121.285	-59.4
1.85	15.53482	44.062	-110.852	-60.3
1.89	14.448	39.75818	-104.441	-61.3
1.93	13.84828	33.97285	-95.8689	-62.7
1.98	13.79782	30.98423	-90.4363	-63.1
2.02	13.06851	28.09427	-82.9409	-63.2
2.07	12.68105	26.57365	-78.6273	-63.2

2.11	11.32548	23.92482	-71.011	-63.3
2.15	10.28566	22.22093	-66.3098	-63.5
2.20	8.79756	19.73523	-59.5346	-63.9
2.24	7.82606	18.03842	-54.6049	-64.1
2.29	6.81794	16.03868	-48.6003	-64.2
2.33	6.05477	14.91225	-44.9937	-64.2
2.37	4.86143	12.84934	-38.8098	-64.4
2.42	3.80771	11.24548	-33.9756	-64.7
2.46	2.96065	9.45372	-29.0932	-65.2
2.51	2.12018	7.90053	-24.9601	-66.0
2.55	1.1355	5.98691	-20.0598	-67.6
2.59	0.4862	4.63371	-16.7937	-69.3
2.64	-0.15398	2.9372	-12.8494	-72.7
2.68	-0.40391	2.0647	-10.852	-75.3
2.73	-0.59401	0.95162	-8.25306	-79.7
2.77	-0.59837	0.35747	-6.68965	-82.2
2.81	-0.5215	-0.23393	-5.03876	-81.5
2.86	-0.4357	-0.47742	-4.1705	-78.3
2.90	-0.30892	-0.66417	-3.23709	-73.0
2.95	-0.21219	-0.72389	-2.71671	-69.4
2.99	-0.07283	-0.72808	-2.14275	-65.1
3.03	0.02914	-0.67376	-1.79046	-62.9
3.08	0.14801	-0.55673	-1.51993	-62.8
3.12	0.22005	-0.45076	-1.39309	-64.1
3.17	0.30065	-0.28089	-1.2909	-67.0
3.21	0.34897	-0.17327	-1.25927	-67.9
3.25	0.40471	-0.00855	-1.23236	-66.9
3.30	0.42253	0.09216	-1.2191	-65.2
3.34	0.42051	0.23289	-1.19589	-62.1
3.39	0.3972	0.31109	-1.17912	-60.5
3.43	0.34859	0.37949	-1.16321	-59.4
3.47	0.30468	0.4029	-1.16532	-59.9
3.52	0.2705	0.40183	-1.18794	-61.3
3.56	0.25575	0.39716	-1.20505	-62.2
3.61	0.23702	0.38704	-1.20264	-63.1
3.65	0.20446	0.37409	-1.15633	-63.6
3.69	0.16573	0.34321	-1.05418	-64.0
3.74	0.13436	0.30878	-1.00369	-65.6
Core TCA9d				
Depth	**X-coordinate**	**Y-coordinate**	**Z-coordinate**	**Inclination**
(m)				(°)
0.00	1.96378	9.453	-42.46376	-72.8
0.03	3.98886	11.99222	-50.05216	-71.1
0.06	6.66757	14.76156	-59.51662	-69.8
0.09	9.57977	17.12246	-69.18684	-69.1
0.13	12.08498	18.89573	-77.69633	-68.8
0.16	14.30822	20.27123	-84.7504	-68.6
0.19	17.14207	21.77856	-93.38657	-68.3
0.22	19.51656	22.847	-100.0503	-68.1
0.25	22.74592	23.9188	-108.1149	-67.9
0.28	25.21428	25.32942	-114.2892	-67.4
0.32	28.2089	27.43815	-120.925	-66.6
0.35	30.93553	29.88897	-126.5521	-65.7
0.38	34.91573	34.25124	-134.64	-64.2
0.41	38.42452	39.36118	-142.6441	-62.8
0.44	43.84207	45.75007	-153.9346	-61.3
0.47	48.04338	49.90419	-162.3316	-60.4
0.50	54.06572	55.52096	-174.8121	-59.5

0.54	58.01633	58.32533	-182.4801	-59.0
0.57	61.75388	61.71273	-189.8257	-58.5
0.60	63.54721	63.553	-192.5564	-58.2
0.63	64.77201	64.8464	-193.8496	-57.8
0.66	65.02955	65.64847	-194.5451	-57.7
0.69	64.53622	66.3838	-195.2866	-57.7
0.73	63.49712	66.73265	-194.9604	-57.8
0.76	59.7465	68.00695	-191.8033	-57.8
0.79	56.81171	67.15015	-186.207	-57.8
0.82	52.73023	64.15935	-176.9983	-57.9
0.85	47.69424	61.15814	-168.9113	-58.4
0.88	42.30344	54.47181	-159.7558	-60.0
0.92	38.50609	49.6312	-154.0816	-61.4
0.95	34.4519	43.26921	-146.5883	-63.2
0.98	32.80135	38.9334	-141.0533	-64.3
1.01	30.78988	36.38409	-135.2171	-64.8
1.04	30.60647	34.48693	-131.862	-65.0
1.07	30.83733	33.20876	-129.4598	-65.0
1.10	31.37699	32.83635	-128.2178	-64.7
1.14	32.51183	32.51249	-126.1354	-64.1
1.17	33.21039	32.5732	-123.3301	-63.4
1.20	33.88636	32.07658	-118.6211	-62.4
1.23	33.69111	31.65323	-115.3394	-62.0
1.26	32.43501	30.05552	-110.7648	-62.1
1.29	31.06234	28.61348	-107.9049	-62.5
1.33	29.84406	26.3252	-105.0876	-63.3
1.36	29.23146	24.04209	-103.3022	-64.1
1.39	28.44412	21.83545	-102.654	-65.2
1.42	28.18674	20.94997	-103.9664	-65.9
1.45	28.63689	20.32468	-106.816	-66.5
1.48	29.83063	20.20908	-110.0352	-66.6
1.51	32.14642	20.29497	-114.4873	-66.4
1.55	34.62478	20.84993	-118.9369	-65.9
1.58	37.13425	20.83558	-123.399	-65.6
1.61	38.52983	20.54913	-125.6383	-65.5
1.64	38.95419	20.01985	-124.6277	-65.2
1.67	38.65097	19.02076	-120.6992	-64.9
1.70	37.81059	17.87567	-115.6457	-64.6
1.74	36.59612	17.28721	-111.7025	-64.6
1.77	34.84413	16.68415	-107.8594	-64.8
1.80	33.34248	15.44243	-103.8749	-65.1
1.83	30.46365	14.39057	-95.49961	-65.2
1.86	27.28123	13.75974	-85.17331	-64.8
1.89	23.64109	12.9459	-73.87638	-64.4
1.93	18.88521	11.51569	-61.08784	-64.5
1.96	14.44835	9.73735	-51.2124	-65.8
1.99	11.47308	8.12042	-44.72011	-67.5
2.02	8.08192	5.49966	-36.70253	-70.6
2.05	6.43924	4.01872	-31.88361	-72.6
2.08	5.06399	2.76571	-26.98645	-74.3
2.11	4.37498	2.00866	-23.69863	-75.1
2.15	3.78633	1.40165	-20.04061	-75.2
2.18	3.45918	1.07598	-17.53432	-74.9
2.21	3.09241	0.92941	-14.3961	-73.7
2.24	2.84842	0.87407	-13.04276	-73.4
2.27	2.45635	0.85948	-10.88684	-72.6
2.30	2.09531	0.84427	-9.24418	-72.3
2.34	1.62368	0.79348	-7.43094	-72.3
2.37	1.2449	0.72201	-6.19831	-73.0

2.40	0.79612	0.62138	-4.99204	-75.0
2.43	0.3647	0.42734	-3.87889	-79.1
2.46	0.04999	0.20144	-3.02068	-84.7
2.49	-0.12102	-0.00566	-2.48475	-86.4
2.52	-0.22101	-0.13656	-1.93229	-80.0
2.56	-0.21923	-0.16492	-1.69309	-77.9
2.59	-0.14356	-0.13642	-1.4835	-79.9
2.62	-0.0612	-0.07982	-1.39326	-84.5
2.65	0.06583	0.00599	-1.33654	-86.3
2.68	0.1671	0.07087	-1.3173	-79.8
2.71	0.29823	0.14817	-1.31383	-71.6
2.75	0.39894	0.21206	-1.32896	-65.9
2.78	0.48438	0.26598	-1.35818	-61.9
2.81	0.53783	0.30712	-1.3887	-59.6
2.84	0.59528	0.36138	-1.43522	-57.4
2.87	0.62742	0.40574	-1.46976	-56.2
2.90	0.65016	0.45661	-1.50094	-55.0
2.94	0.67435	0.47873	-1.52033	-54.3
2.97	0.70882	0.51307	-1.53763	-53.1
3.00	0.73549	0.53462	-1.53994	-52.0
3.03	0.7839	0.5337	-1.53058	-50.7
3.06	0.81797	0.52903	-1.51771	-49.8
3.09	0.8634	0.52267	-1.48857	-48.2
3.12	0.89131	0.50794	-1.46205	-47.3
3.16	0.91716	0.4928	-1.41461	-46.0
3.19	0.93085	0.46874	-1.35823	-44.8
3.22	0.92716	0.44847	-1.29975	-43.9
3.25	0.9083	0.4265	-1.24728	-43.5
3.28	0.85297	0.41903	-1.19023	-43.7
3.31	0.78664	0.40781	-1.15262	-44.7
3.35	0.68297	0.41495	-1.134	-47.1
3.38	0.53874	0.45799	-1.14576	-50.7
3.41	0.28139	0.56972	-1.20697	-54.6
3.44	-0.03437	0.72088	-1.31043	-53.2
3.47	-0.65094	1.03633	-1.57381	-43.8
3.50	-1.10444	1.2176	-1.87906	-40.6
3.53	-1.27699	1.36788	-2.51105	-45.2
3.57	-0.95453	1.37566	-3.10874	-54.2
3.60	-0.13889	1.31474	-4.09186	-66.3
3.63	0.26169	1.26775	-4.39472	-68.2
3.66	0.67953	1.20058	-4.18148	-66.1
3.69	0.80487	1.16862	-3.84566	-63.7
3.72	0.91473	1.117	-3.50083	-61.1
3.76	0.95789	1.07532	-3.59398	-61.9
3.79	1.03209	1.00992	-3.97817	-64.2
3.82	1.18415	0.91701	-4.13676	-64.4
3.85	1.56485	0.73441	-3.64342	-58.1
3.88	1.96326	0.55744	-2.91355	-47.6
3.91	2.35211	0.23834	-1.97739	-32.7

Appendix 8

Measurements of the oxides of Al, Ba, Ca, Fe, K, Mg, Mn, Na, S, Sr, Ti, V, Zn and Zr expressed as ratios of SiO_2, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Ratios have been exaggerated by 100 times.

Upper depth	Lower depth	Al_2O_3/SiO_2	Fe_2O_3/SiO_2	Na_2O_3/SiO_2	MgO/SiO_2	K_2O/SiO_2
(m)	(m)					
0.00	0.02	28.96	13.44	2.81	2.30	3.09
0.02	0.04	29.03	18.83	3.11	2.28	3.15
0.04	0.06	29.01	13.35	2.54	2.36	3.09
0.06	0.08	29.03	14.36	2.96	2.40	3.16
0.08	0.10	29.48	13.76	3.31	2.45	3.25
0.10	0.12	31.02	12.49	2.71	2.49	3.20
0.12	0.14	29.99	12.49	2.83	2.74	3.29
0.14	0.16	30.18	12.45	2.99	2.95	3.30
0.16	0.18	30.48	12.45	3.47	3.11	3.53
0.18	0.20	30.62	11.91	3.01	2.83	3.51
0.20	0.22	30.51	11.51	2.65	2.50	3.45
0.22	0.24	29.86	10.86	2.60	2.26	3.28
0.24	0.26	29.75	10.71	2.76	2.28	3.32
0.26	0.28	31.06	11.64	2.87	2.31	3.40
0.28	0.30	32.31	12.10	2.98	2.38	3.47
0.30	0.32	33.68	11.56	2.13	2.05	3.31
0.32	0.34	31.31	11.00	2.41	2.18	3.45
0.34	0.36	30.53	11.25	2.88	2.34	3.53
0.36	0.38	32.31	12.13	2.89	2.49	3.38
0.38	0.40	31.50	12.33	2.94	2.73	3.58
0.40	0.42	30.71	11.89	3.37	2.68	3.59
0.42	0.44	29.10	10.48	2.99	2.13	3.25
0.44	0.46	29.73	10.80	2.64	1.95	3.14
0.46	0.48	28.04	10.26	2.94	1.89	3.03
0.48	0.50	26.77	9.65	2.73	1.79	2.85
0.50	0.52	26.74	10.16	2.73	1.78	2.84
0.52	0.54	28.44	10.18	2.47	1.67	2.92
0.54	0.56	28.29	9.59	2.73	1.89	3.06
0.56	0.58	28.27	9.87	3.45	2.03	3.18
0.58	0.60	27.79	9.70	3.49	2.07	3.31
0.60	0.62	26.17	8.84	2.77	1.71	2.84
0.62	0.64	24.52	7.97	2.52	1.47	2.66
0.64	0.66	26.54	7.97	2.78	1.49	2.74
0.66	0.68	26.19	8.09	2.66	1.47	2.69
0.68	0.70	25.86	8.10	2.64	1.58	2.73
0.70	0.72	26.43	8.18	2.69	1.59	2.80
0.72	0.74	26.49	8.01	2.54	1.51	2.78
0.74	0.76	29.12	9.70	2.50	1.78	2.93
0.76	0.78	29.15	10.72	2.52	1.94	3.08
0.78	0.80	29.41	9.68	2.73	2.03	3.19
0.80	0.82	29.05	9.58	2.60	1.97	3.15
0.82	0.84	29.35	9.87	2.85	2.04	3.21
0.84	0.86	30.10	9.91	2.58	1.98	3.14
0.86	0.88	29.58	9.45	2.46	1.87	3.03
0.88	0.90	30.65	9.48	2.39	1.95	3.04
0.90	0.92	31.45	9.49	2.81	1.86	2.91

0.92	0.94	26.03	7.28	2.37	1.47	2.72
0.94	0.96	29.01	8.45	2.60	1.77	2.77
0.96	0.98	30.18	9.24	2.05	1.96	2.68
0.98	1.00	30.85	9.69	2.25	2.02	2.76
1.00	1.02	30.79	9.34	2.17	2.04	2.75
1.02	1.04	31.61	9.40	2.15	2.10	2.86
1.04	1.06	32.78	9.60	2.22	2.17	2.93
1.06	1.08	32.58	9.02	1.93	2.06	2.85
1.08	1.10	32.28	9.26	1.98	2.03	2.88
1.10	1.12	32.53	9.40	1.91	2.03	2.87
1.12	1.14	32.71	8.90	2.18	2.07	2.93
1.14	1.16	31.43	8.94	2.06	1.96	2.85
1.16	1.18	30.65	8.32	1.98	1.88	2.74
1.18	1.20	31.27	8.60	2.02	1.96	2.88
1.20	1.22	30.66	7.99	2.42	1.91	3.01
1.22	1.24	30.33	7.91	2.95	1.98	3.22
1.24	1.26	31.68	8.26	2.99	2.19	3.46
1.26	1.28	32.05	7.97	3.15	2.01	3.36
1.28	1.30	31.42	7.20	2.37	1.74	3.21
1.30	1.32	32.22	7.07	2.51	1.80	3.07
1.32	1.34	32.20	7.36	2.45	1.76	3.11
1.34	1.36	32.89	7.17	2.68	1.82	3.23
1.36	1.38	33.60	7.11	2.60	1.84	3.35
1.38	1.40	32.46	6.83	2.71	1.78	3.31
1.40	1.42	31.11	6.67	2.69	1.69	3.14
1.42	1.44	31.61	6.61	2.70	1.64	3.14
1.44	1.46	32.66	6.85	2.14	1.58	2.80
1.46	1.48	31.69	6.60	2.13	1.57	2.64
1.48	1.50	30.92	6.51	2.67	1.61	2.85
1.50	1.52	31.59	6.60	3.65	1.70	3.13
1.52	1.54	31.33	6.55	3.48	1.66	3.09
1.54	1.56	30.60	6.42	2.34	1.61	2.94
1.56	1.58	30.76	6.43	3.64	1.57	3.03
1.58	1.60	31.08	6.58	2.31	1.56	2.93
1.60	1.62	31.83	6.64	3.22	1.66	2.84
1.62	1.64	32.02	6.73	3.30	1.60	2.71
1.64	1.66	32.71	6.96	3.44	1.68	2.82
1.66	1.68	32.36	6.74	3.83	1.64	3.03
1.68	1.70	33.07	6.61	3.06	1.61	3.02
1.70	1.72	32.91	6.70	5.03	1.75	3.17
1.72	1.74	32.42	6.74	2.27	1.61	3.01
1.74	1.76	32.47	6.93	2.61	1.62	2.95
1.76	1.78	32.84	7.05	1.90	1.58	2.71
1.78	1.80	33.43	7.22	2.37	1.54	2.60
1.80	1.82	33.19	7.35	1.64	1.48	2.34
1.82	1.84	33.19	7.34	2.12	1.53	2.37
1.84	1.86	33.26	7.58	1.90	1.55	2.36
1.86	1.88	34.21	7.64	1.88	1.54	2.48
1.88	1.90	34.14	7.73	1.91	1.48	2.40
1.90	1.92	35.06	7.67	1.67	1.46	2.46
1.92	1.94	35.77	7.75	1.83	1.47	2.51
1.94	1.96	35.73	7.69	2.21	1.49	2.55
1.96	1.98	34.39	7.51	2.07	1.47	2.50
1.98	2.00	33.20	7.30	1.94	1.48	2.52

Upper depth (m)	Lower depth (m)	CaO/SiO_2	TiO_2/SiO_2	BaO/SiO_2	V_2O_5/SiO_2	MnO/SiO_2
0.00	0.02	1.75	1.92	0.068	0.046	0.191
0.02	0.04	1.73	1.93	0.071	0.046	0.195
0.04	0.06	1.74	1.92	0.044	0.044	0.180
0.06	0.08	1.77	1.91	0.056	0.044	0.187
0.08	0.10	1.87	1.87	0.047	0.041	0.197
0.10	0.12	1.91	1.88	0.059	0.047	0.183
0.12	0.14	2.17	1.91	0.069	0.046	0.173
0.14	0.16	2.40	1.97	0.058	0.047	0.153
0.16	0.18	2.52	2.02	0.060	0.049	0.143
0.18	0.20	2.18	1.94	0.058	0.045	0.145
0.20	0.22	1.88	1.86	0.051	0.043	0.159
0.22	0.24	1.73	1.81	0.059	0.046	0.133
0.24	0.26	1.71	1.78	0.054	0.044	0.129
0.26	0.28	1.73	1.77	0.054	0.042	0.120
0.28	0.30	1.81	1.73	0.056	0.045	0.118
0.30	0.32	1.23	1.74	0.051	0.045	0.106
0.32	0.34	1.47	1.72	0.053	0.045	0.109
0.34	0.36	1.77	1.69	0.057	0.041	0.113
0.36	0.38	1.60	1.85	0.045	0.047	0.125
0.38	0.40	2.08	1.92	0.062	0.046	0.129
0.40	0.42	2.02	1.86	0.060	0.043	0.121
0.42	0.44	1.51	1.77	0.055	0.040	0.106
0.44	0.46	1.12	1.85	0.047	0.043	0.112
0.46	0.48	1.23	1.81	0.042	0.042	0.113
0.48	0.50	1.31	1.79	0.042	0.042	0.125
0.50	0.52	1.26	1.86	0.035	0.042	0.123
0.52	0.54	1.20	1.80	0.047	0.045	0.122
0.54	0.56	1.31	1.80	0.047	0.041	0.103
0.56	0.58	1.42	1.80	0.053	0.042	0.110
0.58	0.60	1.58	1.71	0.055	0.043	0.102
0.60	0.62	1.18	1.83	0.047	0.038	0.095
0.62	0.64	0.97	1.80	0.047	0.037	0.090
0.64	0.66	0.91	2.01	0.044	0.042	0.086
0.66	0.68	0.92	1.94	0.068	0.041	0.087
0.68	0.70	0.99	1.87	0.053	0.043	0.097
0.70	0.72	1.01	1.95	0.037	0.040	0.097
0.72	0.74	0.94	1.96	0.046	0.041	0.087
0.74	0.76	0.93	2.05	0.033	0.048	0.099
0.76	0.78	1.21	1.83	0.045	0.045	0.110
0.78	0.80	1.34	1.83	0.041	0.044	0.094
0.80	0.82	1.32	1.80	0.050	0.041	0.091
0.82	0.84	1.39	1.81	0.051	0.044	0.093
0.84	0.86	1.24	1.86	0.059	0.048	0.091
0.86	0.88	1.19	1.87	0.053	0.044	0.089
0.88	0.90	1.19	1.91	0.055	0.044	0.086
0.90	0.92	1.04	1.98	0.039	0.047	0.084
0.92	0.94	0.81	1.90	0.048	0.041	0.073
0.94	0.96	1.03	1.97	0.042	0.046	0.084
0.96	0.98	1.25	1.92	0.039	0.047	0.096
0.98	1.00	1.40	1.88	0.035	0.043	0.099
1.00	1.02	1.38	1.88	0.030	0.043	0.097
1.02	1.04	1.42	1.79	0.036	0.045	0.096
1.04	1.06	1.41	1.80	0.038	0.042	0.096
1.06	1.08	1.39	1.77	0.038	0.045	0.100
1.08	1.10	1.37	1.74	0.036	0.044	0.093
1.10	1.12	1.32	1.71	0.038	0.042	0.091

1.12	1.14	1.37	1.70	0.029	0.040	0.088
1.14	1.16	1.41	1.71	0.033	0.044	0.090
1.16	1.18	1.35	1.74	0.033	0.043	0.085
1.18	1.20	1.38	1.74	0.045	0.045	0.086
1.20	1.22	1.39	1.72	0.043	0.043	0.079
1.22	1.24	1.43	1.77	0.063	0.043	0.072
1.24	1.26	1.53	1.87	0.065	0.046	0.072
1.26	1.28	1.36	1.80	0.058	0.045	0.071
1.28	1.30	1.23	1.74	0.048	0.046	0.062
1.30	1.32	1.16	1.72	0.048	0.041	0.059
1.32	1.34	1.17	1.78	0.051	0.044	0.060
1.34	1.36	1.20	1.84	0.066	0.046	0.058
1.36	1.38	1.18	1.92	0.044	0.042	0.053
1.38	1.40	1.13	1.90	0.061	0.045	0.054
1.40	1.42	1.09	1.80	0.053	0.046	0.051
1.42	1.44	1.09	1.75	0.053	0.044	0.050
1.44	1.46	0.99	1.67	0.049	0.043	0.051
1.46	1.48	0.96	1.65	0.038	0.041	0.049
1.48	1.50	1.02	1.65	0.051	0.042	0.047
1.50	1.52	1.10	1.74	0.057	0.043	0.048
1.52	1.54	1.08	1.74	0.055	0.046	0.048
1.54	1.56	1.00	1.71	0.052	0.042	0.049
1.56	1.58	0.98	1.72	0.054	0.040	0.045
1.58	1.60	0.94	1.72	0.055	0.040	0.047
1.60	1.62	0.93	1.77	0.048	0.041	0.048
1.62	1.64	0.89	1.73	0.046	0.037	0.046
1.64	1.66	0.96	1.75	0.051	0.042	0.051
1.66	1.68	1.00	1.77	0.055	0.043	0.048
1.68	1.70	0.99	1.77	0.051	0.041	0.048
1.70	1.72	1.00	1.79	0.054	0.043	0.047
1.72	1.74	0.97	1.79	0.054	0.041	0.048
1.74	1.76	0.97	1.78	0.057	0.041	0.050
1.76	1.78	0.90	1.80	0.051	0.042	0.051
1.78	1.80	0.83	1.82	0.053	0.044	0.049
1.80	1.82	0.81	2.10	0.038	0.044	0.051
1.82	1.84	0.81	1.80	0.046	0.042	0.051
1.84	1.86	0.80	1.82	0.048	0.044	0.050
1.86	1.88	0.74	1.84	0.048	0.044	0.048
1.88	1.90	0.71	1.85	0.043	0.043	0.045
1.90	1.92	0.70	1.87	0.049	0.044	0.046
1.92	1.94	0.71	1.86	0.046	0.042	0.046
1.94	1.96	0.70	1.85	0.045	0.046	0.045
1.96	1.98	0.63	1.86	0.039	0.043	0.037
1.98	2.00	0.60	1.86	0.041	0.041	0.037

Upper depth	Lower depth	ZnO/SiO_2	SrO/SiO_2	ZrO_3/SiO_2	SO_3/SiO_2
(m)	(m)				
0.00	0.02	0.022	0.031	0.036	0.650
0.02	0.04	0.022	0.031	0.037	0.680
0.04	0.06	0.019	0.029	0.034	0.518
0.06	0.08	0.017	0.029	0.033	0.621
0.08	0.10	0.020	0.029	0.033	0.946
0.10	0.12	0.022	0.029	0.031	0.835
0.12	0.14	0.019	0.035	0.033	0.889
0.14	0.16	0.016	0.038	0.035	0.930
0.16	0.18	0.015	0.040	0.036	0.604
0.18	0.20	0.017	0.035	0.035	0.528
0.20	0.22	0.017	0.030	0.032	0.631
0.22	0.24	0.020	0.027	0.033	0.501
0.24	0.26	0.016	0.029	0.033	0.593
0.26	0.28	0.018	0.028	0.031	0.944
0.28	0.30	0.019	0.026	0.030	1.218
0.30	0.32	0.015	0.017	0.030	0.906
0.32	0.34	0.017	0.024	0.032	0.462
0.34	0.36	0.017	0.030	0.031	0.470
0.36	0.38	0.021	0.023	0.031	0.754
0.38	0.40	0.017	0.033	0.035	0.468
0.40	0.42	0.017	0.034	0.034	0.433
0.42	0.44	0.015	0.026	0.035	0.451
0.44	0.46	0.017	0.019	0.034	0.550
0.46	0.48	0.016	0.022	0.033	0.622
0.48	0.50	0.014	0.021	0.037	0.585
0.50	0.52	0.013	0.021	0.033	1.189
0.52	0.54	0.015	0.023	0.037	1.382
0.54	0.56	0.013	0.023	0.032	0.605
0.56	0.58	0.015	0.024	0.033	0.579
0.58	0.60	0.013	0.029	0.034	0.381
0.60	0.62	0.010	0.023	0.038	0.294
0.62	0.64	0.008	0.018	0.037	0.193
0.64	0.66	0.010	0.019	0.036	0.176
0.66	0.68	0.010	0.019	0.037	0.209
0.68	0.70	0.010	0.020	0.034	0.249
0.70	0.72	0.009	0.019	0.033	0.254
0.72	0.74	0.009	0.019	0.034	0.228
0.74	0.76	0.011	0.013	0.033	0.579
0.76	0.78	0.012	0.019	0.031	0.945
0.78	0.80	0.013	0.022	0.031	0.496
0.80	0.82	0.013	0.024	0.032	0.386
0.82	0.84	0.015	0.024	0.031	0.459
0.84	0.86	0.014	0.021	0.030	0.607
0.86	0.88	0.013	0.023	0.036	0.245
0.88	0.90	0.015	0.020	0.033	0.251
0.90	0.92	0.015	0.017	0.034	0.399
0.92	0.94	0.010	0.014	0.034	0.256
0.94	0.96	0.011	0.015	0.032	0.508
0.96	0.98	0.014	0.016	0.031	0.775
0.98	1.00	0.011	0.017	0.028	0.783
1.00	1.02	0.011	0.017	0.028	0.661
1.02	1.04	0.013	0.019	0.028	0.644
1.04	1.06	0.019	0.021	0.027	0.581
1.06	1.08	0.015	0.019	0.028	0.592
1.08	1.10	0.013	0.019	0.027	0.803
1.10	1.12	0.015	0.019	0.025	0.499

1.12	1.14	0.015	0.019	0.027	0.535
1.14	1.16	0.017	0.019	0.029	0.590
1.16	1.18	0.014	0.019	0.029	0.583
1.18	1.20	0.014	0.020	0.027	0.477
1.20	1.22	0.015	0.025	0.031	0.321
1.22	1.24	0.015	0.028	0.031	0.330
1.24	1.26	0.017	0.031	0.030	0.268
1.26	1.28	0.017	0.026	0.030	0.291
1.28	1.30	0.013	0.024	0.028	0.239
1.30	1.32	0.013	0.023	0.029	0.270
1.32	1.34	0.015	0.024	0.027	0.223
1.34	1.36	0.015	0.026	0.029	0.197
1.36	1.38	0.013	0.026	0.033	0.144
1.38	1.40	0.013	0.025	0.030	0.161
1.40	1.42	0.012	0.023	0.030	0.174
1.42	1.44	0.011	0.023	0.028	0.183
1.44	1.46	0.011	0.018	0.027	0.207
1.46	1.48	0.009	0.018	0.027	0.193
1.48	1.50	0.012	0.021	0.026	0.163
1.50	1.52	0.011	0.021	0.028	0.200
1.52	1.54	0.012	0.023	0.028	0.194
1.54	1.56	0.010	0.021	0.028	0.177
1.56	1.58	0.010	0.019	0.038	0.203
1.58	1.60	0.012	0.019	0.028	0.209
1.60	1.62	0.012	0.018	0.030	0.274
1.62	1.64	0.012	0.018	0.027	0.260
1.64	1.66	0.013	0.016	0.029	0.272
1.66	1.68	0.012	0.020	0.030	0.252
1.68	1.70	0.012	0.019	0.032	0.282
1.70	1.72	0.013	0.020	0.031	0.337
1.72	1.74	0.013	0.022	0.030	0.228
1.74	1.76	0.013	0.020	0.034	0.273
1.76	1.78	0.011	0.018	0.033	0.246
1.78	1.80	0.011	0.016	0.029	0.266
1.80	1.82	0.011	0.013	0.029	0.366
1.82	1.84	0.011	0.015	0.028	0.369
1.84	1.86	0.013	0.015	0.030	0.402
1.86	1.88	0.013	0.013	0.031	0.353
1.88	1.90	0.011	0.013	0.029	0.329
1.90	1.92	0.011	0.013	0.031	0.353
1.92	1.94	0.015	0.013	0.033	0.433
1.94	1.96	0.013	0.013	0.030	0.355
1.96	1.98	0.009	0.012	0.032	0.344
1.98	2.00	0.009	0.011	0.032	0.389

Appendix 9

Measurements of P_2O_5 (expressed as ratios of SiO_2, Al_2O_3 and Fe_2O_3), core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Ratios have been exaggerated by 100 times.

Upper depth (m)	Lower depth (m)	P_2O_5/SiO_2	P_2O_5/Al_2O_3	P_2O_5/Fe_2O_3
0.00	0.02	0.62	2.15	4.62
0.02	0.04	0.63	2.16	3.34
0.04	0.06	0.62	2.14	4.65
0.06	0.08	0.65	2.25	4.56
0.08	0.10	0.73	2.48	5.32
0.10	0.12	0.68	2.19	5.44
0.12	0.14	0.61	2.02	4.86
0.14	0.16	0.56	1.85	4.48
0.16	0.18	0.56	1.83	4.47
0.18	0.20	0.53	1.73	4.45
0.20	0.22	0.54	1.77	4.69
0.22	0.24	0.41	1.36	3.74
0.24	0.26	0.39	1.30	3.60
0.26	0.28	0.41	1.31	3.49
0.28	0.30	0.39	1.22	3.26
0.30	0.32	0.34	1.01	2.95
0.32	0.34	0.37	1.18	3.37
0.34	0.36	0.41	1.35	3.67
0.36	0.38	0.46	1.44	3.82
0.38	0.40	0.48	1.51	3.87
0.40	0.42	0.44	1.43	3.70
0.42	0.44	0.36	1.22	3.39
0.44	0.46	0.38	1.27	3.50
0.46	0.48	0.39	1.38	3.79
0.48	0.50	0.39	1.46	4.06
0.50	0.52	0.35	1.32	3.49
0.52	0.54	0.36	1.26	3.51
0.54	0.56	0.32	1.12	3.31
0.56	0.58	0.34	1.22	3.48
0.58	0.60	0.32	1.16	3.31
0.60	0.62	0.33	1.26	3.74
0.62	0.64	0.37	1.50	4.60
0.64	0.66	0.26	0.99	3.30
0.66	0.68	0.27	1.03	3.34
0.68	0.70	0.33	1.29	4.12
0.70	0.72	0.30	1.15	3.72
0.72	0.74	0.25	0.95	3.13
0.74	0.76	0.40	1.38	4.13
0.76	0.78	0.64	2.20	5.97
0.78	0.80	0.41	1.38	4.21
0.80	0.82	0.41	1.42	4.30
0.82	0.84	0.49	1.66	4.93
0.84	0.86	0.47	1.58	4.79
0.86	0.88	0.51	1.73	5.40
0.88	0.90	0.46	1.51	4.88
0.90	0.92	0.41	1.30	4.32

0.92	0.94	0.30	1.16	4.16
0.94	0.96	0.36	1.26	4.32
0.96	0.98	0.43	1.43	4.65
0.98	1.00	0.46	1.48	4.71
1.00	1.02	0.45	1.46	4.81
1.02	1.04	0.45	1.43	4.81
1.04	1.06	0.45	1.37	4.70
1.06	1.08	0.44	1.35	4.89
1.08	1.10	0.44	1.37	4.77
1.10	1.12	0.44	1.34	4.63
1.12	1.14	0.43	1.32	4.86
1.14	1.16	0.43	1.35	4.76
1.16	1.18	0.42	1.36	5.01
1.18	1.20	0.40	1.28	4.65
1.20	1.22	0.36	1.17	4.51
1.22	1.24	0.33	1.10	4.21
1.24	1.26	0.34	1.06	4.08
1.26	1.28	0.32	1.01	4.05
1.28	1.30	0.30	0.94	4.11
1.30	1.32	0.30	0.93	4.25
1.32	1.34	0.29	0.91	3.98
1.34	1.36	0.29	0.88	4.01
1.36	1.38	0.28	0.84	3.98
1.38	1.40	0.27	0.83	3.95
1.40	1.42	0.27	0.86	3.99
1.42	1.44	0.26	0.82	3.93
1.44	1.46	0.27	0.82	3.92
1.46	1.48	0.26	0.83	3.98
1.48	1.50	0.25	0.81	3.86
1.50	1.52	0.29	0.93	4.43
1.52	1.54	0.26	0.83	3.96
1.54	1.56	0.23	0.75	3.57
1.56	1.58	0.25	0.81	3.85
1.58	1.60	0.22	0.72	3.39
1.60	1.62	0.25	0.77	3.70
1.62	1.64	0.25	0.78	3.73
1.64	1.66	0.26	0.79	3.73
1.66	1.68	0.25	0.77	3.71
1.68	1.70	0.24	0.72	3.58
1.70	1.72	0.27	0.83	4.08
1.72	1.74	0.23	0.72	3.48
1.74	1.76	0.24	0.74	3.48
1.76	1.78	0.25	0.77	3.58
1.78	1.80	0.27	0.81	3.75
1.80	1.82	0.26	0.78	3.53
1.82	1.84	0.27	0.80	3.62
1.84	1.86	0.27	0.83	3.63
1.86	1.88	0.25	0.74	3.33
1.88	1.90	0.25	0.72	3.18
1.90	1.92	0.25	0.71	3.24
1.92	1.94	0.25	0.71	3.28
1.94	1.96	0.26	0.73	3.40
1.96	1.98	0.27	0.79	3.60
1.98	2.00	0.28	0.86	3.90

Appendix 10

Measurements of Cd, Cr, Cu and Pb, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. Measures of error represent 95% confidence limits of the measurements.

Upper depth	Lower depth	Cd	error	Cr	error	Cu	error	Pb	error
(m)	(m)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)
0.00	0.02	0.822	0.241	36.413	0.589	21.555	0.314	21.917	1.144
0.02	0.04	0.714	0.209	38.446	0.622	21.179	0.309	22.240	1.161
0.04	0.06	0.741	0.217	37.194	0.602	21.000	0.306	22.015	1.149
0.06	0.08	0.785	0.230	38.680	0.626	23.075	0.336	22.500	1.175
0.08	0.10	0.683	0.200	35.294	0.571	21.485	0.313	19.652	1.026
0.10	0.12	0.646	0.189	35.597	0.576	22.088	0.322	22.310	1.165
0.12	0.14	0.681	0.200	37.819	0.612	22.466	0.327	19.532	1.020
0.14	0.16	0.673	0.197	38.315	0.620	23.389	0.341	16.493	0.861
0.16	0.18	0.492	0.144	39.884	0.645	24.194	0.353	17.633	0.920
0.18	0.20	0.546	0.160	36.527	0.591	22.474	0.328	17.012	0.888
0.20	0.22	0.623	0.182	34.719	0.562	20.915	0.305	18.681	0.975
0.22	0.24	0.570	0.167	34.931	0.565	19.419	0.283	19.375	1.011
0.24	0.26	0.575	0.168	34.278	0.555	19.428	0.283	19.820	1.035
0.26	0.28	0.661	0.194	38.875	0.629	17.959	0.262	18.820	0.982
0.28	0.30	0.499	0.146	37.393	0.605	18.388	0.268	16.931	0.884
0.30	0.32	0.510	0.149	41.431	0.670	14.096	0.205	16.829	0.878
0.32	0.34	0.604	0.177	37.190	0.602	17.339	0.253	16.585	0.866
0.34	0.36	0.571	0.167	34.262	0.554	21.045	0.307	16.697	0.872
0.36	0.38	0.566	0.166	40.323	0.652	19.656	0.287	18.263	0.953
0.38	0.40	0.494	0.145	36.130	0.584	21.139	0.308	15.717	0.820
0.40	0.42	0.633	0.185	36.387	0.589	20.957	0.305	16.144	0.843
0.42	0.44	0.472	0.138	36.818	0.596	19.501	0.284	18.024	0.941
0.44	0.46	0.504	0.148	39.110	0.633	17.647	0.257	19.702	1.028
0.46	0.48	0.968	0.283	36.156	0.585	19.054	0.278	21.184	1.106
0.48	0.50	0.460	0.135	37.140	0.601	19.245	0.281	20.443	1.067
0.50	0.52	0.481	0.141	37.120	0.601	23.198	0.338	20.974	1.095
0.52	0.54	0.516	0.151	36.362	0.588	18.283	0.266	18.484	0.965
0.54	0.56	0.449	0.132	35.724	0.578	18.754	0.273	18.436	0.962
0.56	0.58	0.634	0.186	35.377	0.572	19.031	0.277	18.250	0.953
0.58	0.60	0.464	0.136	31.033	0.502	18.970	0.276	16.592	0.866
0.60	0.62	0.594	0.174	36.210	0.586	20.150	0.294	21.778	1.137
0.62	0.64	0.462	0.135	35.690	0.577	16.908	0.246	23.087	1.205
0.64	0.66	0.465	0.136	36.111	0.584	16.308	0.238	21.450	1.120
0.66	0.68	0.475	0.139	37.479	0.606	16.254	0.237	22.006	1.149
0.68	0.70	0.553	0.162	37.188	0.602	17.415	0.254	21.117	1.102
0.70	0.72	0.363	0.106	37.121	0.601	17.125	0.250	21.272	1.110
0.72	0.74	0.442	0.129	37.678	0.610	17.221	0.251	20.045	1.046
0.74	0.76	0.728	0.213	38.542	0.624	18.831	0.274	26.651	1.391
0.76	0.78	0.649	0.190	35.333	0.572	20.958	0.305	23.006	1.201
0.78	0.80	0.456	0.134	35.542	0.575	24.337	0.355	19.788	1.033
0.80	0.82	0.400	0.117	34.834	0.564	22.587	0.329	21.403	1.117
0.82	0.84	0.702	0.206	34.885	0.564	21.454	0.313	19.887	1.038
0.84	0.86	0.420	0.123	37.150	0.601	21.080	0.307	16.934	0.884
0.86	0.88	0.442	0.129	39.207	0.634	17.821	0.260	17.149	0.895
0.88	0.90	0.344	0.101	37.426	0.605	19.140	0.279	16.983	0.886
0.90	0.92	0.362	0.106	38.623	0.625	18.523	0.270	16.010	0.836

0.92	0.94	0.191	0.056	36.775	0.595	13.588	0.198	14.514	0.758
0.94	0.96	0.745	0.218	37.843	0.612	17.821	0.260	15.779	0.824
0.96	0.98	0.341	0.100	36.334	0.588	20.918	0.305	12.123	0.633
0.98	1.00	0.293	0.086	37.174	0.601	25.944	0.378	12.202	0.637
1.00	1.02	0.392	0.115	35.914	0.581	23.235	0.339	11.385	0.594
1.02	1.04	0.347	0.101	36.832	0.596	22.943	0.334	11.701	0.611
1.04	1.06	0.335	0.098	36.072	0.584	22.707	0.331	11.008	0.575
1.06	1.08	0.412	0.121	35.653	0.577	22.175	0.323	11.584	0.605
1.08	1.10	0.390	0.114	37.569	0.608	23.203	0.338	11.562	0.604
1.10	1.12	0.377	0.110	37.560	0.608	22.461	0.327	12.492	0.652
1.12	1.14	0.160	0.047	35.639	0.577	23.868	0.348	13.589	0.709
1.14	1.16	0.258	0.076	36.268	0.587	23.201	0.338	12.124	0.633
1.16	1.18	0.335	0.098	37.163	0.601	25.442	0.371	10.779	0.563
1.18	1.20	0.267	0.078	38.833	0.628	26.266	0.383	12.170	0.635
1.20	1.22	0.463	0.135	41.919	0.678	27.927	0.407	12.351	0.645
1.22	1.24	0.514	0.151	43.635	0.706	31.567	0.460	14.386	0.751
1.24	1.26	0.326	0.095	42.561	0.689	27.287	0.398	14.416	0.752
1.26	1.28	0.838	0.245	42.355	0.685	28.581	0.417	14.251	0.744
1.28	1.30	0.197	0.058	43.154	0.698	26.597	0.388	12.322	0.643
1.30	1.32	0.309	0.091	44.363	0.718	30.091	0.439	13.546	0.707
1.32	1.34	0.492	0.144	43.503	0.704	28.439	0.415	13.386	0.699
1.34	1.36	0.452	0.132	43.258	0.700	28.462	0.415	14.372	0.750
1.36	1.38	0.504	0.148	43.649	0.706	28.472	0.415	13.409	0.700
1.38	1.40	0.351	0.103	44.169	0.715	28.740	0.419	13.455	0.702
1.40	1.42	0.000	0.000	43.812	0.709	27.453	0.400	14.529	0.758
1.42	1.44	0.298	0.087	44.158	0.714	26.412	0.385	12.273	0.641
1.44	1.46	0.256	0.075	44.900	0.726	27.313	0.398	13.425	0.701
1.46	1.48	0.172	0.050	41.993	0.679	24.481	0.357	10.937	0.571
1.48	1.50	0.293	0.086	40.292	0.652	22.991	0.335	15.003	0.783
1.50	1.52	0.281	0.082	38.911	0.629	23.511	0.343	12.677	0.662
1.52	1.54	0.304	0.089	40.731	0.659	23.132	0.337	14.118	0.737
1.54	1.56	0.207	0.060	40.260	0.651	23.740	0.346	13.336	0.696
1.56	1.58	0.261	0.077	40.907	0.662	24.345	0.355	12.955	0.676
1.58	1.60	0.267	0.078	43.632	0.706	24.288	0.354	14.146	0.738
1.60	1.62	0.175	0.051	43.203	0.699	25.966	0.378	13.000	0.679
1.62	1.64	0.276	0.081	43.279	0.700	27.632	0.403	12.460	0.650
1.64	1.66	0.340	0.100	41.086	0.665	26.431	0.385	15.481	0.808
1.66	1.68	0.480	0.141	40.467	0.655	27.025	0.394	13.762	0.718
1.68	1.70	0.295	0.086	41.969	0.679	25.941	0.378	15.077	0.787
1.70	1.72	0.374	0.109	39.774	0.643	25.778	0.376	13.911	0.726
1.72	1.74	0.263	0.077	41.125	0.665	25.725	0.375	13.648	0.712
1.74	1.76	0.320	0.094	43.460	0.703	26.425	0.385	13.631	0.712
1.76	1.78	0.271	0.079	42.787	0.692	25.179	0.367	13.192	0.689
1.78	1.80	0.488	0.143	44.671	0.723	24.912	0.363	13.726	0.716
1.80	1.82	0.265	0.078	45.704	0.739	26.713	0.389	12.729	0.664
1.82	1.84	0.185	0.054	45.869	0.742	28.081	0.409	14.440	0.754
1.84	1.86	0.315	0.092	44.159	0.714	21.212	0.309	13.888	0.725
1.86	1.88	0.430	0.126	44.463	0.719	23.286	0.339	11.887	0.621
1.88	1.90	0.303	0.089	45.206	0.731	23.677	0.345	12.538	0.654
1.90	1.92	0.866	0.254	43.934	0.711	23.194	0.338	13.709	0.716
1.92	1.94	1.141	0.334	42.499	0.688	21.429	0.312	12.606	0.658
1.94	1.96	0.386	0.113	42.114	0.681	21.595	0.315	14.203	0.741
1.96	1.98	0.331	0.097	40.925	0.662	22.183	0.323	14.688	0.767
1.98	2.00	0.233	0.068	38.795	0.628	22.700	0.331	15.972	0.834

Appendix 11

Unsupported ^{210}Pb activity, core TCA9b, Tocal Homestead Lagoon, Paterson, eastern Australia. The quoted errors of excess ^{210}Pb were determined by a series of standard error propagation calculations which incorporate the uncertainties associated with chemical yield tracers, sample weighing and the spectrometric counting techniques. Activities are normalised to the date of sampling in the field.

Depth (m)	Excess ^{210}Pb	Error
	(Bq kg^{-1})	(Bq kg^{-1})
0.00–0.02	49.13	3.08
0.02–0.04	41.37	3.39
0.06–0.08	50.43	3.35
0.10–0.12	48.26	2.67
0.12–0.14	43.37	3.01
0.14–0.16	18.49	2.17
0.16–0.18	17.88	2.38
0.18–0.20	21.16	3.02
0.22–0.24	18.72	1.90
0.24–0.26	23.62	2.71
0.28–0.30	18.48	1.65
0.30–0.32	14.74	2.25
0.34–0.36	10.98	1.62
0.36–0.38	26.20	2.79
0.38–0.40	11.25	1.95
0.42–0.44	17.55	1.90
0.44–0.46	21.62	1.80
0.48–0.50	10.26	1.98
0.50–0.52	21.43	1.72
0.54–0.56	17.24	2.39
0.56–0.58	10.88	1.50
0.60–0.62	9.18	1.71
0.62–0.64	6.24	1.59
0.66–0.68	5.31	3.10
0.68–0.70	9.95	1.23
0.72–0.74	7.60	3.89
0.74–0.76	9.75	1.63
0.78–0.80	2.47	1.83
0.80–0.82	2.39	1.49
0.84–0.86	3.33	1.94
0.86–0.88	-1.10	-1.85
0.90–0.92	7.49	1.94
0.92–0.94	-1.68	-1.42
0.96–0.98	4.26	1.82
0.98–1.00	5.44	1.27
1.02–1.04	0.67	1.71
1.06–1.08	5.52	1.37
1.08–1.10	2.55	1.35
1.10–1.12	3.53	1.81
1.14–1.16	0.69	1.57
1.18–1.20	5.79	1.44
1.20–1.22	3.14	1.34
1.22–1.24	0.094	2.09
1.24–1.26	0.13	1.60

Appendix 12

Caesium-137 activity, core TCA9b, Tocal Homestead Lagoon, Paterson, central eastern **New South Wales. Measures** of error represent 66% confidence limits of the **measurements. Sample** 0.56-0.54 m (*) was selected for duplicate analysis, neither **duplicate yielded detectable** ^{137}Cs. Sediments below 0.44 m had no detectable ^{137}Cs.

Depth (m)	^{137}Cs (Bq kg^{-1})	Error (Bq kg^{-1})
0.04–0.02	3.7	± 0.3
0.18-0.16	8.5	± 1.0
0.26-0.24	19.0	± 2.2
0.36-0.34	20.5	± 3.4
0.40-0.38	16.7	± 2.0
0.44-0.42	9.5	± 1.0
0.46-0.44	< 8	
0.48-0.46	< 8.7	
0.50-0.48	< 8.7	
0.52-0.50	< 8.7	
0.54-0.52	< 6.9	
0.56–0.54*	< 0.75	
0.58-.0.56	< 1	
0.62-0.60	< 1	
0.64-0.62	< 1	
0.80-0.78	< 4.1	

Appendix 13

Major and minor elemental concentrations of the palaeosol TCWC1, sampled from the catchment of Tocal Homestead Lagoon, Paterson, eastern Australia. The concentration of Al in the sample from 0.01 m depth saturated the spectrometer, and no reading could be made. The code 'bd' denotes those samples (n = 2) that had concentrations below the detection limits.

Mean depth	Ag	Al	Ba	Cd	Co	Cr	Cu
(m)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)
0.01	0.637	s	187.175	0.882	11.618	92.092	27.072
0.04	0.208	36228.609	178.463	0.475	11.076	89.771	20.725
0.07	0.211	35393.632	183.211	0.295	10.276	91.808	20.016
0.10	0.053	29329.289	195.546	0.000	9.461	74.385	16.828
0.13	bd	29283.344	215.927	0.006	12.564	79.508	39.822
0.15	bd	27543.657	262.017	0.000	18.832	76.377	16.084
0.18	0.076	29950.188	172.957	0.000	10.637	87.307	16.461
0.21	0.063	29870.827	168.921	0.078	10.270	86.496	14.779
0.24	0.082	28372.121	167.586	0.000	9.942	81.070	20.508
0.27	0.015	29812.594	176.216	0.000	10.086	88.230	19.098
0.30	0.030	30472.562	196.496	0.063	10.509	89.499	21.014
Mean depth	**Fe**	**Mn**	**Ni**	**Pb**	**V**	**Zn**	
(m)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	($\mu g\ g^{-1}$)	
0.01	34124.883	366.548	39.143	173.267	62.325	180.106	
0.04	35832.035	419.408	37.729	139.456	58.759	117.979	
0.07	35618.710	371.028	39.537	129.635	59.283	114.072	
0.10	30511.592	337.496	33.161	104.680	56.222	65.583	
0.13	34506.482	417.623	35.318	121.549	63.964	65.116	
0.15	34141.417	622.239	36.601	108.381	61.164	56.082	
0.18	36373.529	360.727	39.900	124.283	61.704	58.165	
0.21	36796.576	377.407	39.187	138.037	61.667	62.382	
0.24	36506.893	391.092	42.208	174.409	60.585	65.339	
0.27	38701.900	391.285	41.029	158.507	61.588	78.677	
0.30	37905.121	442.038	41.919	151.761	59.447	95.595	

Appendix 14

Major and minor elemental concentrations of soils sampled from the catchment of Tocal Homestead Lagoon, Paterson, eastern Australia. The code 'bd' denotes a sample which had elemental concentrations below detection limits.

Soil sample	Al	Fe	Mg	Ca	K	Mn	S	P
	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})
T4w	16441.245	22831.113	1440.798	1343.581	1142.642	340.133	794.622	296.204
T17w	24417.056	30234.965	5002.339	6188.392	1947.025	582.374	969.174	657.742
T18	28233.485	36997.349	2433.010	3508.430	1067.688	541.400	1403.953	840.459
T18g	26868.042	37393.434	2897.918	2093.555	1609.186	3381.318	1157.940	491.058
T19d	5609.891	18652.918	363.510	711.961	642.097	139.057	745.317	298.879
T20	9894.807	12071.348	699.301	1526.298	1493.837	182.118	734.863	852.667
T22	7660.692	14277.564	494.935	294.716	317.162	34.401	737.972	97.589
T23a	8079.311	25892.364	540.774	850.717	549.914	120.961	785.379	350.924
T23e	6331.631	13390.589	784.282	952.192	1281.410	124.573	835.312	269.890
T23c	6145.335	14928.054	400.029	427.151	644.538	59.986	771.813	125.314
T24N	8749.202	16209.483	675.018	685.738	686.046	145.120	915.419	184.851
T24S	5719.499	8786.396	784.288	968.468	981.192	133.597	1389.271	133.339
T25	7669.728	11224.561	754.774	1013.763	875.841	102.359	1404.518	477.641
TS4w*	22962.564	28204.929	1615.821	1070.105	1350.464	339.022	777.996	312.274
TS17w	23356.057	30889.967	5039.846	6309.191	2223.334	585.228	548.950	658.765
TS18	37287.649	41341.191	2682.930	3376.213	1296.446	545.162	824.097	834.328
TS18g*	32024.824	38405.903	2921.033	2032.679	1728.136	3115.906	748.218	498.884
TS19d	6457.126	19272.361	413.741	791.142	782.852	140.527	649.102	312.470
TS20	12600.469	14345.244	783.388	1360.493	1645.608	196.116	697.245	891.678
TS22	9564.683	14664.555	553.543	322.149	370.865	38.741	614.442	105.704
TS23a	9616.628	24623.754	619.498	961.782	714.633	130.409	667.057	375.312
TS23c	7648.171	14828.833	462.472	414.585	739.324	61.673	493.180	131.871
TS23e	10011.580	22540.257	836.411	1049.611	1476.569	123.865	599.269	276.199
TS24S	7246.828	10363.934	828.640	935.644	1092.226	144.071	486.702	115.755
TS24N	10604.251	15671.555	746.057	796.650	846.913	148.320	800.792	198.358
TS25	12040.359	13710.302	914.494	1365.880	1309.227	110.485	702.139	529.592

Soil sample	Na	Ba	Zn	Cr	Sr	Pb	Cu	Cd
	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})	(µg g^{-1})
T4w	193.652	104.037	35.309	20.376	22.769	16.868	8.003	0.457
T17w	315.948	91.521	65.626	33.016	73.956	12.897	19.533	0.538
T18	457.022	241.197	59.387	37.156	38.514	21.201	16.348	0.551
T18g	415.953	145.621	55.850	35.484	42.321	26.488	16.846	0.562
T19d	104.594	24.200	18.030	12.773	6.355	11.966	2.749	0.379
T20	152.665	50.766	45.781	12.930	14.264	10.635	4.837	0.273
T22	198.785	25.888	9.684	18.526	6.016	10.816	4.363	0.234
T23a	114.585	36.074	19.719	35.081	5.301	11.771	4.469	0.891
T23e	113.887	32.718	29.447	15.810	nd	10.567	6.306	0.471
T23c	203.597	25.160	14.936	11.008	4.485	9.539	1.941	0.234
T24N	138.724	50.551	19.933	24.544	7.735	9.787	4.193	0.291
T24S	193.755	39.260	23.809	10.042	12.857	10.529	5.522	0.343
T25	123.738	33.739	30.795	10.126	8.258	9.222	6.269	0.365
TS4w*	241.494	108.428	36.271	23.558	24.477	16.107	7.767	0.473
TS17w	344.941	89.408	64.741	32.548	72.781	12.298	16.768	0.380
TS18	483.078	239.570	61.814	40.129	36.316	20.339	17.096	0.605

TS18g*	395.049	144.651	55.102	37.777	42.288	24.600	16.255	0.526
TS19d	99.695	26.720	20.243	13.393	6.786	11.296	3.322	0.456
TS20	130.277	58.210	49.168	15.289	15.528	11.907	8.721	0.213
TS22	229.322	28.744	11.371	22.536	6.553	9.880	3.269	0.153
TS23a	148.816	40.706	22.193	37.623	6.146	11.586	4.500	0.495
TS23c	115.528	27.075	15.107	12.417	4.524	9.390	2.308	0.175
TS23e	140.479	34.750	30.398	17.861	10.342	11.602	6.368	0.382
TS24S	199.726	42.988	24.557	11.909	13.350	12.595	3.978	0.111
TS24N	200.565	53.963	22.910	28.715	8.246	11.507	5.683	bd
TS25	136.459	37.143	32.091	12.963	9.592	9.492	4.524	0.214