Holocene millennial/centennial-scale multiproxy cyclicity in temperate eastern Australian estuary sediments

C. GREGORY SKILBECK,1* TIMOTHY C. ROLPH,2 NATALIE HILL,2 JONATHAN WOODS1 AND ROY H. WILKENS3

1Department of Environmental Sciences, University of Technology, Sydney, Australia
2School of Geosciences, University of Newcastle, Callaghan, Australia
3Hawai‘i Institute of Geophysics, University of Hawai‘i, Honolulu, USA


ABSTRACT: We have undertaken a comparative study of down-core variation in multiproxy palaeoclimate data (magnetic susceptibility, calcium carbonate content and total organic carbon) from two coastal water bodies (Myall and Tuggerah Lakes) in temperate eastern Australia to identify local, regional and global-forcing factors within Holocene estuarine sediments. The two lakes lie within the same temperate climate zone adjacent to the Tasman Sea, but are not part of the same catchment and drain different geological provinces. One is essentially a freshwater coastal lake whereas the other is a brackish back-barrier lagoon. Despite these differences, data from two sites in each of the two lakes have allowed us to investigate and compare cyclicity in otherwise uniform, single facies sediments within the frequency range of 200–2000 years, limited by the sedimentation rate within the lakes and our sample requirements. We have auto- and cross-correlated strong periodicities at 360 years, 500–530 years, 270–290 years, 420–450 years and 210 years, and subordinate periods of 650 years, 1200–1400 years and 1800 years. Our thesis is that climate is the only regionally available mechanism available to control common millennial and centennial scale cyclicality in these sediments, given the geographical and other differences. However, regional climate may not be the dominant effect at any single time and either location. Within the range of frequency spectral peaks we have identified, several fall within known long-term periodical fluctuations of sun spot activity; however, feedback loops associated with short-term orbital variation, such as Dansgaard–Oeschger cycles, and the relationship between these and palaeo-ENSO variation, are also possible contributors. Copyright © 2005 John Wiley & Sons, Ltd.

KEYWORDS: Holocene; multiproxy data; spectral analysis; estuarine sediments; palaeoclimate.

Introduction

The recognition of climate cyclicity in deep marine or non-marine sediments is commonly accompanied by a regular alternation of facies such as the carbonate/clay cycles of deep ocean cores (e.g. Shackelton and Opdyke, 1973; Dean and Gardner, 1985; Diester-Haass and Rothe, 1988; Diester-Haass, 1991), or the regular alternation of evaporite/epiclastites in salt lakes (e.g. Kotwicki and Isdale, 1991) or other ‘closed’ systems (e.g. Horiuchi et al., 2000) that directly or indirectly reflect climate-induced facies variation. In none of these situations do eustatic sea level rise and fall directly influence sediment accumulation, and the vertical sequences therefore represent continuous sedimentation across several Milankovitch cycles, at least in the case of the deep marine sequences. Generally, however, deep-ocean sedimentation rates are insufficient to allow sampling at rates high enough to define centennial or millennial scale cyclicality. Facies alternations also occur in coastal and shallow marine deposits, but as a direct consequence of sea-level rise and fall. In these settings sedimentation is rarely continuous across several facies or sedimentation cycles. For example, where highstand deposits are imbricately stacked, there is little chance of a continuous record spanning more than one sea level cycle being preserved in any one place. This means that sequences in littoral settings usually include non-depositional or erosional breaks at Milankovitch frequencies. However, the coastal zone does provide a good location in which to study cyclicity at sub-Milankovitch

* Correspondence to: C. Gregory Skilbeck, Department of Environmental Sciences, University of Technology, Sydney, PO Box 123, Broadway, NSW 2007, Australia. E-mail: g.skilbeck@uts.edu.au
In order to assess the effects of external or regional controls on sedimentation, we have analysed the sedimentary records of two New South Wales (Australia) coastal lakes (Myall and Tuggerah Lakes, Fig. 1). Both are located inboard of the Holocene highstand beach system and, on the basis of the $^{14}$C age sequence and correlatable high-resolution magnetic susceptibility records, contain a stratigraphically continuous Holocene estuarine sequence overlying either an earlier Pleistocene erosion surface, or lowstand late Pleistocene fluvial deposits. The estuarine sediments were deposited as sea level rose and drowned the coastal areas following the last glacial maximum, some 20 000 calibrated years ago. In this paper, we describe and compare downhole variation in three parameters common to our two studies, magnetic susceptibility, percentage calcium carbonate ($\%$CaCO$_3$) and percentage total organic carbon ($\%$TOC) from the Holocene estuary sediments in the two lakes. Temporal variation in these properties will reflect the changing nature of the local environment, providing a signal of the ecosystem response to direct, or indirect (e.g. sea-level change) climate forcing functions. Time series analysis of these properties demonstrates the potential for deciphering local and regional controls on sedimentation and, potentially, on climate in temperate eastern Australia over at least the last 10 000 years.

Lake settings

The southeastern Australian coast is a wave-dominated sediment-deficient stable passive margin (Roy and Boyd, 1996) that formed during opening of the Tasman Sea 80–55 million years ago (Weissel and Hayes, 1977). The Myall Lakes System (which includes Myall, Broadwater and Boolambayte lakes, Fig. 2) overlies irregular Carboniferous (~320 Ma) basement comprising rhyodacitic-to-basaltic forearc basin volcanics and metasediments of the New England Fold Belt (Skilbeck and Cawood, 1994). To the north and west, basement rocks crop out around the lakeshores, and most of the small islands within the lakes comprise basement outcrops. None of the cores to date have intersected rocky basement and the pattern of depth to bedrock is essentially unknown. On the seaward side of the lakes, the Pleistocene and Holocene dunes (Melville, 1984; and Boyd, 1996, Fig. 2) link headlands formed of Carboniferous basement outcrop. The minimum topographic relief between Myall Lake and the adjacent ocean is 20 m above mean sea level, meaning that this part of the system is essentially isolated from direct marine influence. The maximum water depth approaches 5 m, although lake level is known to fluctuate up to 80 cm above sea level (D. Rissik, pers. comm., 2001), mainly as a result of rainwater influx, but at equilibrium approximates local sea level. The lake system has an indirect marine connection at Port Stephens, some 30 km to the southwest of the lake system (Fig. 1). Despite this connection, and its proximity to the sea, Myall Lake contains virtually fresh water (2–3 ppt TDS) and has no existing tidal or external wave-current influences. This situation is unique along the New South Wales coast where all other lakes, including Tuggerah Lake, are either directly or periodically open to the sea and contain widespread reworked marine sand deposits.

Tuggerah Lake (part of the Tuggerah, Munmorah and Budgewoi Lakes system) is a barrier estuary (Roy, 1984) formed within a valley incised into the Triassic Narrabean strata of the foreland Sydney Basin (Glen and Beckett, 1997). Tuggerah Lake has a maximum water depth of 3 m and fully saline to brackish waters. It is semi-enclosed by a coastal sand barrier but is in permanent communication with the Tasman Sea through a microtidal inlet (The Entrance) located near the southeastern end of the lake. Quaternary sediments are extremely variable in thickness having been deposited within and adjacent to at least two incised channel systems, during multiple phases of sea level rise and fall (Weale, 2001).

The two lakes therefore have some attributes in common (inter alia regional setting; area, water depth, geomorphology, Tertiary history) and some that differ (inter alia provenance; current marine influence). It is relevant to our study that the rivers feeding the two lakes drain distinctly different geological provinces (Fig. 1), but because some of the units in the Sydney Basin were derived from erosion of the New England Fold Belt (Hamilton and Galloway, 1989), a common lithological provenance cannot be excluded when trying to assess regional and local controls on sedimentation.

Sediment description

We have recovered 36 cores from Myall Lakes (referred to herein as ML#) and 2 from Tuggerah Lake (Pelican 1 and Chittaway 1), using a combination of vibrocoring (in 75 mm diameter aluminium liner), push-piston and hammer coring methods (range of 32 mm to 90 mm diameter plastic liner).
Figure 2  Map and cross sections from Myall Lakes showing facies distribution in selected cores. Note that in the north of the lake (ML13, 12, 24) that Holocene highstand estuary deposits overlie an erosion surface beneath which probable MIS 5e orange mottled estuarine clay subcrops. Downhole logs adjacent to stratigraphic sections are low-frequency magnetic susceptibility (in cgs × 10^{-6} units). For facies key refer to Fig. 4.
Penetration ranges from 0.45 m (ML29, by vibrocoring) to 10.77 m (ML34, push-piston core) in Myall Lakes, and up to 4.34 m (Pelican-1, hammer core) in Tuggerah Lake. Although currently undated, the oldest sediments encountered in cores in both lakes are interpreted to be highstand estuary deposits emplaced during the last interglacial highstand (MIS 5e) and subaerially exposed during the last glacial maximum (MIS 2). These sediments are very similar in appearance to the Holocene estuarine muds described below, but are considerably stiffer. The uppermost parts of these units have orange-brown iron oxide mottles.

In the deeper basinal parts of both lakes the main facies is a pale–medium grey (5GY/3; Munsell, 1975) silty clay containing mostly disseminated grains of fine to medium-grained subangular-rounded quartz sand and irregularly distributed fragmentary and rare intact bivalve shells (Fig. 3). This unit is interpreted to be a highstand central basin estuarine facies, that, where depositionally complete, ranges in thickness up to 1.74 m in Tuggerah Lakes (Pelican-1), and up to 2.05 m (ML07) in Myall Lakes. The facies is uniform in appearance, although rare dark grey mottling may indicate some localised bioturbation. Internal bedding is rare; in a few cores (ML14 and 19) sand is concentrated into thin laminae near the top of the unit, and a 10 cm sandy silt layer is present in Pelican-1 in Tuggerah Lake. Common shell fragments and rarer whole

<table>
<thead>
<tr>
<th>Core</th>
<th>Proxy</th>
<th>Age min. (yr)</th>
<th>Age max. (yr)</th>
<th>Range (yr)</th>
<th>T_{\text{max}} (yr)</th>
<th>T_{\text{min}} (yr)</th>
<th>N</th>
<th>Core type</th>
<th>Sample</th>
<th>Core diameter (mm)</th>
<th>Core spacing (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Myall 19A</td>
<td>Magnetic susceptibility</td>
<td>1603</td>
<td>8175</td>
<td>6572</td>
<td>2196</td>
<td>(1.3)</td>
<td>107</td>
<td>V</td>
<td>1.3 a</td>
<td>75</td>
<td>3.5 a</td>
</tr>
<tr>
<td>Myall 11B</td>
<td>Magnetic susceptibility</td>
<td>1965</td>
<td>8175</td>
<td>6200</td>
<td>2640</td>
<td>(1.3)</td>
<td>94</td>
<td>V</td>
<td>1.3 a</td>
<td>75</td>
<td>3.5 a</td>
</tr>
<tr>
<td></td>
<td>TOC</td>
<td>1965</td>
<td>8175</td>
<td>6200</td>
<td>2640</td>
<td>(1.3)</td>
<td>94</td>
<td>V</td>
<td>2.6 a</td>
<td>58</td>
<td>2.6 a</td>
</tr>
<tr>
<td></td>
<td>CaCO_3</td>
<td>1965</td>
<td>8175</td>
<td>6200</td>
<td>2640</td>
<td>(1.3)</td>
<td>94</td>
<td>V</td>
<td>2.6 a</td>
<td>58</td>
<td>2.6 a</td>
</tr>
<tr>
<td>Myall 19B</td>
<td>Magnetic susceptibility</td>
<td>7560</td>
<td>12508</td>
<td>5028</td>
<td>4270</td>
<td>(1.3)</td>
<td>1200</td>
<td>P</td>
<td>0.5</td>
<td>32</td>
<td>2.6 a</td>
</tr>
<tr>
<td></td>
<td>TOC</td>
<td>7560</td>
<td>12508</td>
<td>5028</td>
<td>4270</td>
<td>(1.3)</td>
<td>1200</td>
<td>P</td>
<td>3</td>
<td>32</td>
<td>2.6 a</td>
</tr>
<tr>
<td></td>
<td>CaCO_3</td>
<td>7560</td>
<td>12508</td>
<td>5028</td>
<td>4270</td>
<td>(1.3)</td>
<td>1200</td>
<td>P</td>
<td>3</td>
<td>32</td>
<td>2.6 a</td>
</tr>
<tr>
<td>Pelican 1</td>
<td>Magnetic susceptibility</td>
<td>3355</td>
<td>7704</td>
<td>4373</td>
<td>1820</td>
<td>(2.6)</td>
<td>81</td>
<td>P</td>
<td>2</td>
<td>35</td>
<td>2.6 a</td>
</tr>
<tr>
<td></td>
<td>TOC</td>
<td>3355</td>
<td>7704</td>
<td>4373</td>
<td>1820</td>
<td>(2.6)</td>
<td>81</td>
<td>P</td>
<td>3</td>
<td>35</td>
<td>2.6 a</td>
</tr>
<tr>
<td></td>
<td>CaCO_3</td>
<td>3355</td>
<td>7704</td>
<td>4373</td>
<td>1820</td>
<td>(2.6)</td>
<td>81</td>
<td>P</td>
<td>3</td>
<td>35</td>
<td>2.6 a</td>
</tr>
<tr>
<td>Chittaway 1</td>
<td>Magnetic susceptibility</td>
<td>2022</td>
<td>5002</td>
<td>2980</td>
<td>2022</td>
<td>(2.6)</td>
<td>108</td>
<td>P</td>
<td>2</td>
<td>35</td>
<td>2.6 a</td>
</tr>
<tr>
<td></td>
<td>TOC</td>
<td>2022</td>
<td>5002</td>
<td>2980</td>
<td>2022</td>
<td>(2.6)</td>
<td>108</td>
<td>P</td>
<td>3</td>
<td>35</td>
<td>2.6 a</td>
</tr>
<tr>
<td></td>
<td>CaCO_3</td>
<td>2022</td>
<td>5002</td>
<td>2980</td>
<td>2022</td>
<td>(2.6)</td>
<td>108</td>
<td>P</td>
<td>3</td>
<td>35</td>
<td>2.6 a</td>
</tr>
</tbody>
</table>

*Uncompacted (1 cm) and 2 cm respectively in core. T_{\text{max}} and T_{\text{min}} are the range for which reliable frequency peaks can be determined given by SPECTRUM analysis; core type: V = vibrocoring; P = piston core, H = hammer core.

*To maximum age of 10 k 14C yr BP.
Figure 4  Downhole proxy data for Tuggerah and Myall Lakes; (a) Pelican 1, (b) Chittaway 1, (c) ML11B, (d) ML19A and (e) ML19B. Magnetic susceptibility data in all cores are low-frequency volume measurements. In Pelican 1 magnetic susceptibility reached a maximum of 443 cgs x 10^-6 units near the base of the core. Ages in ML11B and ML19A are cal. yr BP with ±2σ range (95% confidence). Correlation tie point of 0 m in ML19B shown on Fig. 4(d); age tie points from ML19A shown on Fig. 4(e).
shells (Figs 3 and 4), dominantly of the bivalves *Anadara* sp. and *Notospisula trigonella*, are present in many of the cores in the middle part of the estuarine unit. Shell concentrations within this zone produce layers up to 15 cm thick (e.g. Chittaway-1), while in places the shell material appears to be present in two or three poorly defined bands (Figs 3 and 4) mostly up to a few cm thick. Elsewhere shells and shell fragments are distributed irregularly and lack a preferred orientation. Magnetic susceptibility data, supported by 14C dates, indicate that shell beds/layers do not correlate in age. Minor components are variably scattered throughout the estuarine facies, and include charophyte gyrogonites (in the upper part of the unit, immediately beneath the gyttja facies), common authigenic pyrite, and irregularly distributed wood and charcoal fragments.

Along the landward side of both lakes, a combination of sandy silt or silty sand is the uppermost sediment present, in beds up to 30 cm thick. This unit represents progradation of fluvial clastic sediments (bayhead deltas; e.g. Chittaway-1, Fig. 3). The sediments comprise mainly lithic silty sand, with variable amounts of mud and organic material. Along the seaward margin of both, well-sorted fine to coarse-grained quartz sand and silty sand, represents either flood tidal deltas (e.g. Pelican-1, Fig. 1) and/or aeolian dune migration (e.g. ML02, ML26, Fig. 2). In both cases, the coarse-grained sand units are up to 3 m thick and contain shell material, fibrous plant remains and charcoal. Well-defined vertical and horizontal burrows are variably present. In all cases, the coarse-grained beds are restricted to the margins of the lake. Nowhere have either fluvial or marine sand bodies migrated completely over finer-grained estuarine sediments in the central part of either lake.

In the central parts of Myall Lake, a layer of olive-yellow/green amorphous organic matter (AOM), up to 1.6 m thick, overlays the grey silty estuarine clay. This sediment, or gyttja, is soupy (in the upper 40–60 cm) to gelatinous (down-core) in consistency. Its base has been dated at around 1110 ± 140 cal. yr BP (OZD298, Table 2). It is the youngest unit intersected in all cores away from the edges of Myall Lake. In
all cases the boundary between the gyttja and the underlying estuarine clay is gradational over 10–20 cm. Minor components include disseminated quartz and lithic sand-grains, relatively abundant charophyte remains, and subordinate black, well-rounded faecal pellets. Much of the floor of the central part of Myall Lakes is covered by weeds dominated by the macroalgae *Najas marina* (prickly waternymph) and it is thought that the breakdown of this material has contributed most of the organic mass of the gyttja facies. Although this unit is Late Holocene in age, we have excluded it from our analysis because of highly variable sedimentation rates calculated at different sites, mainly a result of highly variable amounts of compaction.

Underlying the estuarine clay one of two facies types are present:

1 Around the margins of Myall Lakes, and in the two Tuggerah cores, the underlying facies is a grey silty clay of similar appearance and composition to the Holocene estuarine sediment described above, but with prominent orange–reddish brown mottling, and a much stiffer consistency. Where this sediment is present, the boundary is invariably sharp, and probably erosional. In Pelican-1 the immediately overlying facies is a coarse, intra-formational lag in which the clasts are composed of angular oxidised clay pellets clearly derived from the underlying material. In Chittaway-1 and in all Myall Lakes cores, however, the lag deposit is absent and the younger, softer estuarine clay immediately overlies the stiffer unit. We interpret this underlying unit as an estuarine, central basin facies, probably accumulated during the MIS 5e highstand, in environments similar to those existing today. The unconformity and reddish-brown staining indicate subaerial exposure and probable erosion that we believe occurred during the intervening MIS 5d-2 period, prior to the last postglacial marine transgression.

2 In the central parts of Myall Lakes, the underlying facies is a dark brown or grey to black, organic-rich (TOC up to 24%, ML19) structureless silty clay, or sapropel. It has common to abundant disseminated plant and woody material, much of which is coated with iron monosulphides (Fig. 2). Disseminated quartz and lithic grains occur near the base of the unit. Gypsum crystals and unidentified sponge spicules of at least two types occur irregularly throughout. Rare, thin oxidised horizons (e.g. ML11A) suggest periodic subaerial exposure of the facies. The upper boundary varies from sharp (e.g. ML01, 09, and 32) to gradational over 20–70 cm (ML03, 11A,B, 22, 28) or mottled and bioturbated (ML07, 19, 20, 21). We interpret this unit to represent overbank deposition in a semi-permanent fluvial standing water body such as a swamp, during the lowstand conditions that would have dominated the area from MIS 5d-2.
<table>
<thead>
<tr>
<th>Laboratory code/ Core depth (cm)</th>
<th>Core depth (cm)</th>
<th>Uncompacted core depth (cm)</th>
<th>14C age (radiocarbon years)</th>
<th>± 1σ error (cal. yr BP)</th>
<th>Calibrated age (1σ)</th>
<th>± 1σ error (cal. yr)</th>
<th>Rounded ± 1σ error (cal. yr)</th>
<th>% area enclosed by 1σ probability distribution</th>
<th>Rounded ± 2σ error (cal. yr)</th>
<th>% area enclosed by 2σ probability distribution</th>
<th>δ13C (per mil)</th>
<th>% modern C</th>
<th>Sample type</th>
<th>Calibration curve</th>
<th>ΔR (cal. yr)</th>
<th>ΔR ± 1σ error (cal. yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>OZD298 ML19A 40</td>
<td>54</td>
<td>1200.0</td>
<td>70.3</td>
<td>1110</td>
<td>60</td>
<td>0.815</td>
<td>140</td>
<td>1.000</td>
<td>22.3</td>
<td>86.4</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD299 ML19A 60</td>
<td>81</td>
<td>2228.2</td>
<td>60.8</td>
<td>2210</td>
<td>60</td>
<td>0.790</td>
<td>120</td>
<td>0.990</td>
<td>23.5</td>
<td>75.9</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD300 ML19A 140</td>
<td>188</td>
<td>81.462</td>
<td>96.2</td>
<td>90.60</td>
<td>80</td>
<td>0.649</td>
<td>250</td>
<td>0.887</td>
<td>23.7</td>
<td>36.3</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD301 ML19A 160</td>
<td>215</td>
<td>88.320</td>
<td>75.7</td>
<td>98.90</td>
<td>100</td>
<td>0.658</td>
<td>260</td>
<td>0.985</td>
<td>24.8</td>
<td>33.3</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD302 ML19A 180</td>
<td>242</td>
<td>83.512</td>
<td>78.2</td>
<td>93.60</td>
<td>60</td>
<td>0.584</td>
<td>130</td>
<td>0.854</td>
<td>27.6</td>
<td>35.3</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD430 ML11B 100</td>
<td>118</td>
<td>63.44</td>
<td>40.7</td>
<td>580</td>
<td>20</td>
<td>0.664</td>
<td>50</td>
<td>1.000</td>
<td>18.4</td>
<td>93.0</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD431 ML11B 118</td>
<td>138</td>
<td>1357.4</td>
<td>40.9</td>
<td>1280</td>
<td>20</td>
<td>0.931</td>
<td>60</td>
<td>0.889</td>
<td>22.4</td>
<td>84.7</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD434 ML11B 245</td>
<td>288</td>
<td>7706.7</td>
<td>55.0</td>
<td>8480</td>
<td>50</td>
<td>0.911</td>
<td>90</td>
<td>1.000</td>
<td>28.6</td>
<td>38.2</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD435 ML11B 284</td>
<td>335</td>
<td>9076.2</td>
<td>57.7</td>
<td>10220</td>
<td>40</td>
<td>0.935</td>
<td>140</td>
<td>0.990</td>
<td>28.1</td>
<td>32.2</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD432 ML11B 170</td>
<td>200</td>
<td>4312.6</td>
<td>39.0</td>
<td>44.20</td>
<td>130</td>
<td>1.000</td>
<td>280</td>
<td>1.000</td>
<td>0.04</td>
<td>58.5</td>
<td>S</td>
<td>M98</td>
<td>7</td>
<td>86</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OZD433 ML11B 196</td>
<td>231</td>
<td>5561.7</td>
<td>39.4</td>
<td>5940</td>
<td>120</td>
<td>1.000</td>
<td>220</td>
<td>1.000</td>
<td>0.19</td>
<td>49.9</td>
<td>S</td>
<td>M98</td>
<td>7</td>
<td>86</td>
<td></td>
<td></td>
</tr>
<tr>
<td>WK8741 P1 231</td>
<td>231</td>
<td>3170.0</td>
<td>250.0</td>
<td>3380</td>
<td>280</td>
<td>0.928</td>
<td>580</td>
<td>0.992</td>
<td>24.4</td>
<td>67.4</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>WK8742 P1 363</td>
<td>363</td>
<td>5840.0</td>
<td>330.0</td>
<td>6660</td>
<td>360</td>
<td>0.980</td>
<td>670</td>
<td>0.986</td>
<td>23.8</td>
<td>48.3</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>WK8743 P1 406</td>
<td>406</td>
<td>8130.0</td>
<td>160.0</td>
<td>9030</td>
<td>180</td>
<td>0.784</td>
<td>400</td>
<td>0.983</td>
<td>25.2</td>
<td>36.3</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>WK8744 C1 108</td>
<td>108</td>
<td>1860.0</td>
<td>230.0</td>
<td>1820</td>
<td>260</td>
<td>1.000</td>
<td>500</td>
<td>1.000</td>
<td>26.9</td>
<td>79.3</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>WK8745 C1 235</td>
<td>235</td>
<td>5300.0</td>
<td>210.0</td>
<td>6090</td>
<td>200</td>
<td>0.954</td>
<td>430</td>
<td>1.000</td>
<td>24.1</td>
<td>51.7</td>
<td>B</td>
<td>198</td>
<td>—</td>
<td>—</td>
<td></td>
<td></td>
</tr>
<tr>
<td>WK10406 C1 328</td>
<td>328</td>
<td>3545.0</td>
<td>101.0</td>
<td>3550</td>
<td>150</td>
<td>1.000</td>
<td>320</td>
<td>1.000</td>
<td>0.04</td>
<td>63.5</td>
<td>S</td>
<td>M98</td>
<td>7</td>
<td>86</td>
<td></td>
<td></td>
</tr>
<tr>
<td>WK10407 C1 229</td>
<td>229</td>
<td>39480.7</td>
<td>71.0</td>
<td>3910</td>
<td>150</td>
<td>1.000</td>
<td>300</td>
<td>1.000</td>
<td>0.05</td>
<td>61.2</td>
<td>S</td>
<td>M98</td>
<td>7</td>
<td>86</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*OZ = ANSTO AMS Laboratory, Lucas Heights, NSW, Australia; WK = Waikato 14C laboratory, NZ.

*Calibrated age rounded according to Stuiver and Polach (1977).

*B = bulk sediment sample; S = shell.

*bI98 = INTCAL98; M98 = MARINE98.
Lake stratigraphy

The Late Quaternary stratigraphy of Myall Lakes is relatively well-known because of the number of cores available. Tuggerah Lakes has been extensively drilled as part of a coal exploration programme, but unconsolidated sediment, including the Late Quaternary, was only spot-sampled from drilling fluids during the exploration program. The two cores from the current study remain the only continuous lithological samples from the Late Quaternary in Tuggerah Lakes. A study of the sequence stratigraphy in Tuggerah Lakes from a 2D shallow seismic survey was recently undertaken (Weale, 2001).

The stratigraphy of Myall Lakes is summarised in Fig. 4. It can be seen that uppermost gyttja is present commonly across the central, deeper parts of the lake where it is up to 1.63 m thick (ML11B). Close to the margins of the lake, the uppermost facies is sand, which along the southern margin is well sorted and quartz-rich and represents the landward edge of the relict/modern dune complex. Core ML05 has penetrated through this sandy facies, showing that the sand represents the prograded distal fringes of the Pleistocene/Holocene dune complex and that it is underlain by finer-grained facies. Along the northern margin of the lake, the sand is poorly sorted and lithic, indicating its origin as a prograding fluvial wedge derived from the underlying Carboniferous sedimentary rocks. The volumetrically dominant facies in the lake is the estuarine grey silty clay that is present in all cores, except ML13 located at the northern shoreline of the lake. In the central parts of the lake, where it is gradationally underlain by the sapropel facies and gradationally overlain by gyttja; this estuarine facies is consistently between 1.60 and 2.05 m thick. Magnetic susceptibility data (Fig. 2) suggest deposition has not been uniform across the central part of the lake, despite the uniform appearance of the facies, and the correlatable pattern indicates either that there has been little or no bioturbation, or that the magnetic signature is post-depositional in nature.

The strike of the palaeosubstrate, as indicated by the facies distribution described above, is approximately northeast–southwest, parallel to the current long axis of the lake. Marine and estuarine facies have clearly prograded towards the north during recent evolution, but the absence of recent estuarine clay in ML13 suggests the lake has never extended farther to the north than its current limit.

Dating and sedimentation rates

Chronologies and sediment accumulation rates have been established using conventional (Tuggerah) and AMS (Myall) radiocarbon dating. The conventional analyses were carried out by the Waikato Radiocarbon Dating Laboratory, while the AMS ages were obtained from the Australian Nuclear Science and Technology Organisation (ANSTO) ANTARES facility (Lawson et al., 2000). Results are given in Table 2. The radiocarbon samples from the Tuggerah Lake cores were approximately one-quarter core segments, 4 cm thick, providing between 50 and 100 g of dry weight; the Myall Lakes samples were approximately 1 cm³ in size over a 1 cm interval.

All ages cited herein have been converted to cal. yr BP where possible (i.e. for ages <20000 cal. yr BP), using CALIB Version 4.3 (Stuiver and Reimer, 1993; Stuiver et al., 1998), and have been rounded according to the convention outlined in Stuiver and Polach (1977). Ages are cited in the text and on Fig. 4 with the 2σ error determined from the probability distribution (Method B in CALIB 4.3; see Table 2). The percentage probability distribution associated with both the 1σ and 2σ errors by this method are also given in Table 2. The carbon in bulk sediment samples derives from a mixture of terrestrial and estuarine plant and freshwater algal material and hence the 14C ages were corrected using the atmospheric correction curve IntCal98. The shell samples from Tuggerah Lake were corrected using marine calibration curve MARINE98, because this system is open to the sea. Although Myall Lake is currently a freshwater system, the bivalve fauna is estuarine in character and the marine calibration curve MARINE98 has been used, with the eastern Australian regional reservoir correction (Table 2). For all Holocene AMS ages, the δ13C values have been measured (range −28.59 to −22.33% for bulk sediment samples and −1.89 to −0.44% for shells).

For the Tuggerah Lakes cores, age and proxy samples have been obtained from the same core at each of the two locations, whereas in Myall Lakes ages were obtained from cores ML11B and ML19A and then correlated with adjacent cores from which the proxy samples were collected (ML11A and ML19B respectively) using high-resolution magnetic susceptibility records and calculated sedimentation rates. The age sample locations and tie points are shown on Fig. 4. The method is comparable with that used by Moy et al. (2002) for their study of palaeo-ENSO in the Andes.

Eighteen 14C ages are available from the Holocene estuarine facies in four cores. The downhole plots of calibrated age against uncorrected core depth are shown in Fig. 5. In both Myall and Tuggerah Lakes, the ages determined on bulk sediment samples define consistent downhole sedimentation rates for estuarine sediment where up to four ages are available in one core (ML11A, ML19A; Fig. 5). In Myall Lake, the rates range between 11 and 22 cm per thousand years (kyr), whereas in Tuggerah Lake they are slightly higher at ~30 cm kyr⁻¹. From Fig. 5 it is clear that the bulk sediment samples define consistent downhole sedimentation rates across the Holocene interval, with high correlation coefficients for the linear regression, but shell-derived ages plot well off these regression lines. For the Myall Lake samples, the shells yield ages about 400 yr older than their stratigraphic position suggests compared with the bulk sediment ages, whereas in the Tuggerah cores the ages are between 1500 and 2200 years younger than their stratigraphic position within the sequence of terrestrial dates would indicate. These differences suggest that the average eastern Australian reservoir correction of ΔR = 7 ± 86 yr, is in error by at least 200 yr, but also that a general reservoir correction cannot be applied to littoral faunas. We consider the shell ages to be unreliable and they have not been used for interpolating ages for the spectral analysis.

Of the remaining 14 bulk sediment-derived ages, only two require further discussion. The lowermost date of 9030 ± 400 cal. yr BP (WK8743) at a depth of 4.06 m in Pelican 1 (Fig. 4) is of concern. The bulk sample was taken from the lag layer at the very base of the Holocene section, which contains clay rip-up clasts and thus could be contaminated by older material. The sea level on eastern Australia (Thom and Roy, 1985) indicates that sea level was between ~6 and 9 m below present sea level at this time so this age would yield an index point that lies above the sea-level curve. In addition, shell and wood material dated between ~6800 and 8200 cal. yr BP in nearby Lake Macquarie and the Hunter River estuaries, are present at depths of 9.7–16 m b.s.l. (Thom et al., 1992; Thom et al., 1992; Roy, 1994; Walker, 1999). The younger terrestrial ages in the Pelican-1 and Chittaway-1 cores are consistent with the data of Thom and Roy (1985) indicating that present sea level was reached at ~6500 cal. yr BP. Because of the uncertainties about this older Tuggerah date, we have ignored it for
the purposes of the frequency analysis and used instead an age extrapolated from the two overlying Holocene dates. These yield an extrapolated basal age for the Holocene estuarine sediments of $\sim 7730$ cal. yr BP and a sedimentation rate of $40$ cm kyr$^{-1}$ across the Holocene interval, marginally higher than that calculated from the linear regression shown on Fig. 5.

In ML19A, the lowermost Holocene ages of 9890±260 cal. yr BP (OZD301) and 9360±130 cal. yr BP (OZD302), both lie within the upper part of the sapropel layer underlying the estuarine clay facies and are out of chronostratigraphic order relative to each other. The older age yields a sedimentation rate of $32$ cm kyr$^{-1}$ for the lowermost part of the Holocene section, and is supported by the three other ages across the Holocene section. In contrast, the younger of the two out-of-sequence ages yields a sedimentation rate of $154$ cm kyr$^{-1}$ for the same interval. As the lower of the two rates rate falls well within the range determined by Roy (1994) for Holocene estuarine sedimentation in eastern Australia, and is supported by all other ages determined in this study, the 9360±130 cal. yr BP age has been disregarded for interpolation of multiproxy data.

Figure 5  Depth versus calibrated age (in cal. yr before 1950) showing regression fits for ages in (a) Myall Lake and (b) Tuggerah Lake. Linear regression and Pearson correlation coefficient from MS Excel 2002. See text for explanation of shell-derived ages

Analytical methods

The main factors that can cause changes in down-core sediment characteristics are: (1) changes in the nature of the material entering the water body; (2) authigenic production of sediment components; and (3) alteration of sediment components within the water body or after deposition. These factors will respond on a regional basis directly, or indirectly, to
climate change; the crux of the interpretive procedure is to identify the driving mechanism of a particular sediment parameter. Part of this process is to analyse the power spectrum of the various data and identify common spectral peaks that can be associated with known climate cycles.

The data sets common to both lakes are percentages of total organic carbon (TOC), calcium carbonate and water content and (volume) magnetic susceptibility ($C_{20}$).

**Total organic carbon–calcium carbonate determinations**

In Tuggerah Lakes, TOC and $% C_{a} C_{a}$ were determined using the method of Dean (1974) by % wt LOI at 550°C and 1000°C respectively, after evaporating free interstitial water at 105°C for 12 hours. Organic matter content was converted to organic carbon using a multiplication factor of 0.58, the average range of the weight fraction of carbon in dry peat and lignite (Maher, 1998). In the high resolution Myall samples, these parameters were measured using a Rosemount Dohrmann catalytic combustion chamber in which $C_{a} C_{a}$ was determined by difference between TOC and total carbon after the removal of carbonate by acid digestion. TOC and carbonate data determined from LOI are also available for ML19A. The sample spacing of 10 cm does not yield useful results in the 200–2000 yr band of interest in the frequency analysis, so these have not been incorporated into our spectral study, but the results do confirm the downhole trends in the higher resolution ML11 data (Fig. 4).

**Magnetic susceptibility**

In three of the four cores (Pelican-1, Chittaway-1 and ML11B) volume magnetic susceptibility was measured using a Bartington MS2B dual-frequency sensor attached to a Bartington MS2 meter, and the low-frequency data were used for spectral analysis. Data for core ML19B were measured on a Geotek multi-sensor core logger using an MS2C loop attached to a Bartington MS2 meter. All other magnetic susceptibility data were collected using a high-resolution MS2E surface scanning sensor attached to a Bartington MS2 meter. In both of the latter cases, the values reported are volume determinations measured at a low frequency and are directly comparable with the low-frequency readings of the dual-frequency sensor. Values of magnetic susceptibility in both lakes are reported in cgs units and are uniformly low, except where iron oxide staining is visible in the cores. In order to confirm that the downhole magnetic susceptibility trends shown in Fig. 4 are real, we conducted multiple runs at varying sample spacings (10, 2 and 1 cm) using both high (10 second count, 0.05 cgs units) and low (1 second count, 0.5 cgs units) resolution measurement settings.

In addition to the proxies in common, magnetic parameters ARM (anhysteretic remanent magnetisation) and SIRM (saturation isothermal remanent magnetisation) were obtained for the Tuggerah Lake cores and ML11B, whereas $C_{a} C_{a}$, density and major and trace element geochemistry were obtained from selected Myall Lake cores, including ML19B. Further discussion and spectral analysis is limited to the common parameters %TOC, %CaCo and volume magnetic susceptibility.
Sediment properties (proxies) and driving mechanisms

The organic carbon content of the sediments will be related to the type/source of organic matter and its rate of accumulation (both absolute and with respect to the remainder of the sediment load) and degradation. In estuaries and lakes, the organic matter load is a combination of allochthonous organic matter brought in by rivers and autochthonous production within the water body. The former is dominated by more refractory terrestrial organic matter comprising material from vascular plants that is at least partially oxidised. Conversely, autochthonous organic matter comprises both vascular material (e.g. bottom-rooted emergent plants), and more labile non-vascular (e.g. algae) material. Mineralisation of the organic matter is accomplished through processes mediated by aerobic and anaerobic bacteria; the efficacy of mineralisation depending on the proportion of refractory to labile components and relative importance of the aerobic to anaerobic pathways (Westrich and Berner, 1984; Hedges et al., 1988; Canfield et al., 1993; Canfield, 1994; Kristensen et al., 1995; Hedges and Kiel, 1995; Hulthe et al., 1998). The relative importance of the anaerobic pathway will depend on the exposure time of the organic matter to aerobic conditions before transport below the oxic/anoxic boundary (redoxcline). Typically the redoxcline is located just below the sediment–water interface, so exposure time to aerobic conditions will relate to the water depth, rate of burial, resuspension effects and bioturbation (Canfield, 1994; Aller, 1994; Meyers and Ishiwatari, 1995; Kristensen, 2000). The carbonate content of the sediments will largely reflect the abundance of intact and fragmentary micro- and macrofossil shell material. Other mechanisms that could affect the carbonate content include photosynthesis in the surface water, which can deplete CO₂ levels to the point where calcite supersaturation occurs and calcium carbonate is precipitated (Kelts and Hsu, 1978), and the post-depositional diagenetic dissolution and precipitation of carbonate minerals within the sediment (Berner, 1980).

The quantity, particle-size and type(s) of magnetic minerals entering the water body will determine the initial magnetic properties of the sediments. Changes in the magnetic input may occur through variable catchment erosion and/or changes in the magnetic properties of catchment soils due, for example, to fire (Rummery et al., 1979) or to variations in pedogenic intensity (Tite and Linington, 1975). Subsequent modification within the subaqueous sediments may occur through microbiially mediated reductive diagenesis, authigenesis and biomineralisation (Stoltz et al., 1986; Canfield and Berner, 1987; Karlin et al., 1987; Karlin, 1990; Tarduno and Wilkison, 1996). Based on the ARM data, reductive diagenesis has probably modified the magnetic minerals in the Tuggerah Lakes cores as indicated by the jump in magnetic susceptibility at a depth of ~180 cm (Fig. 4, Chittaway 1). Mineralisation of organic matter in the anoxic sediment has led to the dissolution and pyritisation of primary magnetic minerals and the formation of secondary magnetic minerals. This alteration potentially obscures the primary depositional signal replacing it with a signal related to the intensity of suboxic or sulphidic diagenesis. Consequently, the magnetic data from the Tuggerah cores provides, at least partly, a record of environmental processes that postdate the deposition of the sediment in which that change is
recorded. Spectral analysis of the Tuggerah magnetic data must therefore be viewed with caution.

### Spectral analysis

We have carried out auto- and cross-correlation spectral analysis of the interpolated time series for volume magnetic susceptibility, total organic carbon and carbonate spectral data for each of the two sites in the two lakes studied (Pelican 1, Chittaway 1 from Tuggerah Lake; ML11A/B and ML19A/B from Myall Lake). Temporal record length and sample spacing for each site, and the analytical parameters these affect, are listed in Table 1. Because of the limitations inherent in the study, the combination of site location (which affects the type of facies present) and the sample spacing (which controls the temporal resolution of the data), we have only been able to investigate spectral content in the range 200–2000 yr for the data available. We have employed two different Fourier transform analytical methods, (i) the Blackman–Tukey method (BT), undertaken using the wave analysis software package Igor Pro® from Wavemetrics, and (ii) the Lomb–Scargle method, using the freeware SPECTRUM package (Schultz and Stattegger, 1997). The latter method has the advantage of not requiring evenly- (for a single time series), and evenly- and equally-spaced time data (for pairs of data sets in cross-spectral analysis). Interpolation of data to produce this condition leads to a predictable decline in amplitude at successively higher frequencies (‘red noise’) according to Schultz and Stattegger (1997; see inter alia this paper, and Xanthakis et al. (1995), for discussion and comparison of various spectral analysis methods as applied to palaeoclimate records). The results from SPECTRUM include an estimation of the reliable frequency range, the maximum period for which analysis identifies at least two cycles within each analytical segment of the data set (i.e. the low-frequency limit, $T_{\text{max}}$), and an average (because the data are not equally spaced) Nyquist frequency (i.e. the minimum resolution, $T_{\text{Nyq}}$). On the other hand, IgorPro provides considerably more control over analytical parameters, and in particular, the smoothing algorithms which are used

---

**Figure 8** Autocorrelation (above) and cross-correlation (below) power spectra for magnetic susceptibility data for the cores indicated. Note that ML19A has been included as an extended time series for ML19B. ‘P1’ = Pelican 1; ‘C1’ = Chittaway 1. The x-axis is log$_{10}$ of the period (yr). Asterisked time series are plotted against the right axis; double-asterisked curves are plotted against the far right scale in the upper diagram. All other time series results are plotted against the left axis. Indexed peaks are in yr.
to remove high-amplitude, low-frequency spectral content, beyond the reliable frequency range. In at least some of our data, these are introduced because of boundary effects at the upper and lower stratigraphic contacts of the estuarine clay facies. In most of our data, the sedimentation rate (i.e. the thickness of the Holocene sequence) and dependent sampling frequency have resulted in time series with less than 100 data points. This is less than ideal for the statistical estimations, particularly at levels of confidence above 90%, which accompany individual spectral analysis. We suggest, however, that the use of the multiproxy data and four different sites, and the use of two different analytical methods, allows us a robust cross-check of the significance of the results.

Our study comprised spectral analysis of (i) autocorrelated time series for each available parameter at each site; (ii) cross-correlation of the time series for each of the three different parameters at each coring site; and (iii) cross-correlation of the time series of each common parameter at all sites, using ‘raw’ data in SPECTRUM. The analytical parameters, which were common throughout, are given in Table 3. The results are shown diagrammatically in Figs 6 to 10 which show stacked normalised power spectra for the above combinations of data, converted to \( \log_{10} \) age (period) space, rather than the more common frequency space, for ease of visual inspection. We then carried out spectral analysis of the same data sets using IgorPro to remove or reduce the low-frequency content of the data beyond the 2000 yr limit. Although this operation

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>OSFC</td>
<td>4.0</td>
</tr>
<tr>
<td>HIFAC</td>
<td>1.0</td>
</tr>
<tr>
<td>Segments with 50% overlap</td>
<td>1.0</td>
</tr>
<tr>
<td>Window type</td>
<td>Hanning [3]</td>
</tr>
<tr>
<td>Level of significance</td>
<td>0.10 [2]</td>
</tr>
<tr>
<td>Subtract trend from data segments</td>
<td>Y</td>
</tr>
<tr>
<td>Use logarithmic dB scale</td>
<td>N</td>
</tr>
<tr>
<td>Unit of time</td>
<td>Ka [2]</td>
</tr>
<tr>
<td>Maximum frequency to plot</td>
<td>Nyquist ( \times ) HIFAC [1]</td>
</tr>
</tbody>
</table>

![Figure 9](image-url)  
Autocorrelation (above) and cross-correlation (below) power spectra for CaCO\(_3\) data for the cores indicated. ‘P1’ = Pelican 1; ‘C1’ = Chittaway 1. The x-axis is \( \log_{10} \) of the period (yr). Asterisked time series are plotted against the right axis; all other time series results are plotted against the left axis. Indexed peaks are in yr.
can be carried out automatically in SPECTRUM (via the harmonics menu), IgorPro offers more control over the analysis. Removal of low-frequency information was achieved by applying a 15-point Gaussian smoothing window to the raw data, subtracting these smoothed values from the raw input data, and applying a Fourier transform to the residual curve. The procedure is shown graphically for ML11B carbonate data, together with the resulting variation in the power spectra, in Fig. 11, and the power spectra for the autocorrelation analysis for the time series from the remaining three cores in Fig. 12.

Modelling of ideal data sets containing up to four weighted and unweighted sine waves with both SPECTRUM and IgorPro has shown that the relative amplitude of spectral power peaks is a function of both the relative amplitudes of the various frequencies in the original data (and the number of actual points defining these), as well as the combination of parameters used for the analysis (such the number of points in the analytical segments). In order to minimise the potential effect of spectral amplitude not being tied to peak significance in terms of its geological importance, we have additionally undertaken a simple frequency distribution of all unweighted spectral peaks from all of the SPECTRUM-analysed data sets. The resulting frequency histogram is plotted in Fig. 13. While this procedure weighs the data at the higher end of the frequency spectrum more heavily than at the lower end, it still allows us to identify the most common frequencies as peaks in various parts of the overall 200–2000 yr range.

The standard fast Fourier transform algorithm in IgorPro requires a minimum of between three and five cycles to be present in the raw time series, in order produce reliable determination of their recurrence period. While this limitation is supposed to be partly reduced by the Lomb–Scargle method (to the two cycles defined by the reliable frequency range), we have still chosen to restrict our analysis to the period range 200–2000 yr discussed above. The other potentially critical factor which has not yet been incorporated in our analysis, is the condition of stationarity in the time-series data. We are aware that non-stationarity is present in the magnetic susceptibility data from Pelican-1 in Tuggerah Lake, probably because of diagenetic alteration, and as yet it cannot be excluded from other data sets. However, overall the procedure has still yielded useful results.

Figure 10 Autocorrelation (above) and cross-correlation (below) power spectra for TOC data for the cores indicated. ‘P1’ = Pelican 1; ‘C1’ = Chittaway 1. The x-axis is log10 of the period (yr). Asterisked time series are plotted against the right axis; double-asterisked plots in the lower diagram are plotted against the far right axis. All other time series results are plotted against the left axis. Indexed peaks are in yr.
In our analysis of the results we are most concerned with frequencies that are common between sites and proxies, rather than in the dominant frequencies. Inspection of the results of the SPECTRUM analyses (Figs 6 to 10) has enabled us to identify between four and six quasi-frequency/period peaks which are relatively consistent across the majority of our multiproxy data sets. These are, in chronological order, at $T_{210}$ yr, $T_{250}$ yr, $T_{350-370}$ yr, $T_{420}$ yr, $T_{500-530}$ yr and $T_{1250-1450}$ years. One additional, but less clearly defined frequency, at $T_{1800}$ yr, shows up at the low end of the study range. In identifying these peaks we have relied not only on the actual position (frequency) of the peak, but also on the pattern of the power spectra for the time series. This similarity in spectral pattern but actual frequency mismatch is not unexpected because of the relatively few age tie points available, and resulting small-scale variations in sedimentation rates that could not be accounted for. Overall, the repeatability of the spectral patterns gives us confidence that the data are real. Interestingly, there are no clearly common or coherent frequencies in the range 530–1000 yr. A number of peaks in this range probably represent local cyclicity in various individual parameters.

The longest record available to us at the highest resolution for all proxy data was obtained from ML11B, in the eastern part of Myall Lake (time series range 6160 yr [1965–8125 yr]; $n = 118$ for magnetic susceptibility and $n = 58$ for TOC and CaCO$_3$; $T_{min} = 94$ yr, $T_{max} = 2640$ yr for magnetic susceptibility). Not surprisingly, the highest coherency was obtained for auto- and cross-correlation of the records from this core within the 200–2000 yr analytical range. We identified strong common peaks at all of the frequencies mentioned above from the auto- and cross-spectra for the multiproxy time series (Figs 6 and 7), and one additional peak at around 1000 yr which is strongly present in the carbonate data and as a slight shoulder peak in the carbonate–magnetic susceptibility cross correlation data.

Auto- and cross-spectral analyses of the same proxy data from the other sites (Figs 6 and 7) support the commonality of the mid-300 yr (in the range 340–370 yr) peak, with the exception of the data from ML19B (where the record only covers the early part of the Holocene) and the magnetic susceptibility data from Chittaway-1 (which is affected by diageneric alteration, but has a slight shoulder in the mid-300 yr range). However, a periodicity of $T_{360}$ yr is subordinately present in the

![Figure 11](image-url)
Figure 12 Left: Plots of the residual time series data from (a) ML11B, (b) ML19B, (c) Pelican 1 and (d) Chittaway 1 for each of the data sets indicated. Magnetic susceptibility plotted against the left axis in each case, carbonate plotted against the near right axis and TOC data plotted against the far right axis. Right: Power spectra for each of the adjacent time series, plotted with the x-axis as log10 of the period (yr). The period of the cyclicity represented by the peaks are labelled (yr). Analysis from IgorPro
magnetic susceptibility record of ML19A which spans in excess of 8000 yr of the Holocene. This peak is the also one of three that are common to all residual data sets (Fig. 12).

The ML11B data contain three peaks in the 200–300 yr range ($C_{24}^{210}$, 240–260, 275–280 yr) with the $C_{24}^{80}$ yr peak being the strongest, and the only one present in all three proxy data sets. Most data sets from other sites have one, or more commonly two, peaks in this range, and in all cases it is the mid-200 yr peak which is present. The exception is the TOC and CaCO$_3$ data in Chittaway-1 where the Nyquist frequency ($T_{\text{min}}$ on Table 1) is 271 yr.

There are indications of two or three peak periods in the range 1000–2000 yr in most of the data sets. The most common of these are in the ranges between 1000 and 1100 yr, and 1300–1500 yr. In both cases peaks are present at all sites in at least one data set, although in some (e.g. a minimal shoulder in the Pelican-1 carbonate data), they are suppressed in the BT data from Igor analysis. The 1300–1500-yr period is strongest in the longest time series (ML11B and ML19A), and in the carbonate and TOC cross-correlation data. It is not present in any of the magnetic susceptibility cross-correlation, which is a manifestation of the lowish $T_{\text{max}}$ values in Pelican and Chittaway cores (Table 1).

Discussion

The presence of periodic signals that have an intra- and inter-lake commonality must indicate a driving mechanism, or mechanisms, with at the very least a regional expression. Palaeoenvironmental research in the last two decades has identified variability in the Holocene climate at millennial, and shorter, periods (for reviews see Adams et al., 1999; Chambers et al., 1999; Overpeck and Webb, 2000). Suggested driving mechanisms include solar cycles (periodicities of $C_{24}^{2200}$, $C_{24}^{210}$ (Suess cycles), $C_{24}^{80}$ (Gleissberg cycles), 22 and 11 (Schwabe cycles) yr; e.g. Chambers et al., 1999; Perry and Hsü, 2000) and thermohaline circulation (Dansgaard–Oeschger (D/O) cycles; $C_{24}^{1500}$ yr quasi-periodicity; e.g. Bond et al., 1997). Periodicities of 779 yr and 206 yr in monsoon rainfall intensity have been resolved in the $\delta^{18}O$ record obtained from a stalagmite collected from Oman (Neff et al., 2001) and a 770-yr periodicity is present in marine sea surface salinity records in the South China Sea (Wang et al., 1999) and at ODP Site 658 (Kutzbach and Guetter, 1986) where it is attributed to Asian and African monsoon activity respectively. North Atlantic sediments have yielded periodicities in sediment colour (a proxy for North Atlantic deep water circulation) centred at 550 yr and 1000 yr (Chapman and Shackleton, 2000). Plant cellulose from peat deposits in northeastern China has significant $\delta^{18}O$ periodicity at around 207, 245, 311, 590, 820 and 1046 yr (Hong et al., 2000), while organic matter content in a Scottish lake shows a quasi-periodicity at $C_{24}^{200}$–225 yr (Battarbee et al., 2001). A compilation of peat bog studies from northwest Europe (Chambers and Blackford, 2001) lists centennial-scale cycles with lengths of 200–210, 260, 360, 450, 520 and 800 yr. One possible explanation for the non-obvious expression of the $C_{24}^{1500}$-yr cycle is the minimal amplitude expression of this phenomenon during interglacials,
as is recorded in the GISP2 ice-core signature over the interval 550–5000 yr BP (Schulz et al., 1999), but it could also be a manifestation of the variable frequency of Dansgaard–Oeschger cycles, that range between 1000 and 2000 yr over the Late Pleistocene (MIS 2, 3 and 4).

Of particular interest to temperate and tropical eastern Australia is the long-term variability (and existence) of the El Niño–Southern Oscillation (ENSO) climate phenomenon. Stott et al. (2002) have concluded from a study of Mg/Ca and δ18O in the western Pacific warm pool that a ‘super’ ENSO oscillating in phase with stadial/interstadial (i.e. D/O cycles) has operated in the Pacific region over the past 70 kyr, and a similar conclusion was reached by Moy et al. (2002) from a study of Holocene Andean lake sediments. Many workers have found that ENSO had a reduced intensity over the period from 15 000 to about 6 000 cal yr BP (Sandweiss et al., 1996; Clement et al., 1999; Rittenour et al., 2000; Koutavas et al., 2002; Moy et al., 2002). Bond et al. (2001) proposed a solar forcing mechanism for at least the Holocene part of the North Atlantic 1500-yr cycle, although more recently Schulz and Paul (2002) have called the existence of a 1400–1500-yr North Atlantic cyclicity into question for the Holocene.

Although there are clearly no shortage of candidates with which to match the periodicities obtained from the present study, we are left with essentially uniform, massive Holocene estuarine deposits in at least two coastal lake/lagoon systems, containing at least three proxies which yield common periodicities matching global or semi-global forcing factors at centennial and millennial scale (for instance, solar or deep ocean thermosteric oscillations). Whilst we are unable to prove a direct cause and effect relationship between the variables and the climate factors, and are unable to elucidate relationships such as coherency or lag between these at this stage, we must speculate that global climate factors are influencing the accumulation of one or more components of the sediment system along the NSW coast. It may be that the common periodicities seen in our data and the proxy monsoon records, Chinese peat cellulose accumulations and long-term ENSO variability support the suggestion that trans-equatorial air streams driven by the monsoon and trade winds can link the climate of the two hemispheres (An, 2000).

Our most intriguing result is that the most consistent cyclicity in our records, the 340–360-yr period, is present in the pollen-cout data from the northwest European peat bogs (Wijmstra et al., 1984; Oliver et al., 1997), where it is attributed by Wijmstra et al. to sun-spot cycles. Whether our data reflect a local response to solar variability is uncertain, although a link between solar variability and the strength of ENSO impacts has been suggested for Australia and North America (Franks, 2001).

Comparison of our data with existing Australian records is hampered by a lack of published information on Holocene cyclicity at millennial and centennial scales. The Holocene climate of southeast Australia has been the subject of two recent studies (Dodson, 1998; Schulmeister, 1999). There is a consensus on the rapid amelioration of climate between ~11 500 and 8500 cal yr BP followed by a climatic optimum, at least in effective precipitation, between 7000 and 5000 cal yr BP. The last 5000 yr appears more equivocal, with Dodson (1998) inferring a cooler drier climate, but with an increase in effective precipitation after 3000 cal yr BP, while Schulmeister (1999) infers a sharp reduction in effective precipitation after 5000 cal yr BP. However, there is a dearth of published information regarding Holocene climate cyclicity for the region. The penultimate cold period seen in the northern hemisphere (~3100–2400 yr; e.g. O’Brien et al., 1995) coincides with changes in peat accumulation at Barrington Tops, NSW (Dodson, 1987), southeast Australian lake levels (Bowler and Wasson, 1984), and a ~1-m sea level fall at Port Hacking, NSW (Baker and Haworth, 2000), although at least some of the coastal sea level variation can be attributed to hydroisostasy. However, there needs to be a concerted effort to generate Holocene climate records from the Australian continent that have sufficient resolution and chronological control to address inter-hemispheric climate links and the question of the global nature of Holocene climate cyclicity (e.g. Charles et al., 1996; Broecker, 2001). This will be the target of our future investigations of southeast Australian Holocene sediments.

Conclusions

Our data suggest it is possible to identify meaningful periodicities in down-core data sets where there are less than an ideal number of discrete data points for frequency analysis, using a multiproxy approach. We have demonstrated strong Holocene global climate periodicities within our data over the analytical range 200–2000 yr at ~210 yr, ~250 yr, 350–370 yr, ~420 yr, ~500–530 yr and 1250–1450 yr, and possibly at around 1800 yr. All of these periods have been identified in other palaeoclimate studies of stratigraphic data where they have manifested in both deep marine and non-marine environments and can be attributed to variations in solar insolation or perturbations in orbital cycles brought about by feedback loops.

Our study shows that even in otherwise uniform, single facies sediments, it is possible to measure and meaningfully analyse down-core variation in relatively easy-to-measure physical and chemical parameters. In order to pursue the study to decade scale resolution, and to allow us to address the statistical significance of individual spectral analyses, we must either seek Holocene sections in excess of 6 m thick for bi-decadal resolution, or undertake multiple hole coring at our current sites. The former option is the more attractive because it does not involve the need for detailed and accurate correlation.

Acknowledgements We would like to thank the Australian Nuclear Science and Technology Organisation (ANSTO) through the Australian Institute of Nuclear Science and Engineering (AINSE) granting scheme (grants number 97/195R and 99/124) for the provision of carbon-14 dating. The research has been supported by the Australian Research Council (Small Grant 97/311), Brian Jones and Colin Murray-Wallace provided comments on the original manuscript.

References


