Lake George revisited: New evidence for the origin and evolution of a large closed lake, Southern Tablelands, NSW, Australia

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Lake George contains the longest continuous sedimentary record of any Australian lake basin, but previous age models are equivocal, particularly for the oldest (pre-Quaternary) part of the record. We have applied a combination of cosmogenic nuclide burial dating, magnetostratigraphy and biostratigraphy to determine the age of the basal (fluvial) unit in the basin, the Gearys Gap Formation. Within the differing resolutions achievable by the three dating techniques, our results show that (i) the Gearys Gap Formation, began accumulating at \( \approx 4 \) Ma, in the early Pliocene (Zanclean), and (ii) deposition had ceased by \( \approx 3 \) Ma, in the mid late Pliocene (Piacenzian). Whether the same age control provides an early Pliocene (Zanclean) age for the formation of the lake basin is uncertain.

During the Piacenzian, the vegetation at the core site was a wetland community dominated by members of the coral fern family Gleicheniaceae, while the surrounding dryland vegetation was a mix of sclerophyll and temperate rainforest communities, with the latter including trees and shrubs now endemic to New Guinea—New Caledonia and Tasmania. Mean annual rainfall and temperatures are inferred to have been \( \sim 2000–3000 \) mm, although probably not uniformly distributed throughout the year, and within the mesothermal range (\( >14 ^\circ \)C \( <20 ^\circ \)C), respectively. Unresolved issues are: (1) Does the basal gravel unit predate uplift of the Lake George Range and therefore provide evidence that one of the proposed paleo-spillways of Lake George, that above Geary’s Gap, has been elevated up to 100–200 m by neotectonic activity over the past 4 million years? (2) Did a shallow to deepwater lake exist elsewhere in the lake basin during the Pliocene?

KEY WORDS: Australia, Lake George basin, Lake George Range, Gearys Gap Formation, Neogene, Zanclean, Piacenzian, cosmogenic nuclide burial age, paleomagnetism, palynostratigraphy, Nothofagus temperate rainforest.

INTRODUCTION

Lake George (Indigenous name Weereewaa), which occupies a tectonic depression on the Southern Tablelands of New South Wales at an elevation of \( \sim 675 \) m, is distinguished from other large inland lakes in mainland Australia by its highland setting. Because of its prominent location adjacent to the highway connecting Canberra to Sydney (Figure 1), as well as its dramatic filling and drying episodes (including total desiccation during prolonged droughts), a range of urban myths and legends have been attributed to the lake (references in Barrow 2012). Scientifically testable claims include its longevity, upwards of 5 million years, and that drill core recovered from the lake floor could provide the oldest continuous record of hydrologic changes in Australia and possibly the world (Barrow 2012). The inference of great age is based on a suggested late Miocene age (Truswell 1984, 1985) for microflora preserved in organic horizons between 108.1 and 110.4 m in Core hole C354, one of several mud-rotary holes (Figure 2) drilled by the Bureau of Mineral Resources (now Geoscience Australia) through the then-dry floor of Lake George in 1982/83. The holes were drilled in order to determine the age of the faulting that impounded the lake basin and also to monitor water level and quality in the lake (see Driessen 1974; Abell 1991, pp. 90–91). In contrast, paleomagnetic evidence indicates that the same thin interval of carbonaceous sand and claystone is late Pliocene in age (McEwan Mason 1991).

This paper is the first of several re-evaluating the origins of the Lake George basin and the evolution of Lake George using new evidence from core samples taken...
from BMR Core hole C354 (Figure 1), archived at Geoscience Australia. Here, the focus is on the basal of the three major sedimentary units infilling the Lake George basin, the fluvial Gearys Gap Formation. The specific aims are to (i) improve the age control for the Gearys Gap Formation and thereby establish the maximum age for lacustrine deposition within the lake basin and (ii) review the paleoenvironmental implications of fossil pollen, spores and other plant microfossils preserved in one of the rare sections of unweathered lake facies in
Core hole C354, in order to reconstruct environments potentially preceding the formation of paleo-Lake George. Subsequent papers will discuss the earliest Pleistocene (Gelasian) and late Pleistocene history of the lake, centred on microflora preserved in: (a) the overlying Ondyong Point Formation in BMR Core hole C354, (b) 50–40 ka deposits infilling paleochannels exposed in sand quarries on the eastern side of Lake George, and (c) a reassessment of ANU Core holes LG1-4, drilled close to the site of BMR Core hole C354 (Figure 1) and which provides a record for the past 250 ka (Singh & Geissler 1985).

PHYSICAL SETTING AND GEOLOGIC BACKGROUND

The Southern Tablelands are part of the predominantly north–south chain of mountain ranges, monadnocks and plateaux, with deep gorges, which are collectively known as the eastern highlands or Great Dividing Range and extend some 3500 km from the northeastern tip of Queensland southwards into western Victoria. Lake George (35°08′S, 28°30′E, 675 m elevation) is an ephemeral, north–south orientated lake located ~30 km north-east of Canberra (Figure 1) on the Southern Tablelands between the Blue Mountains west of Sydney and the Southern Alps on the NSW–Victoria border. When filled, the lake is up to 24 km long and 10 km wide. In the Lake George region, the crest of the Great Dividing Range forms the watershed between rivers that flow eastwards into the Pacific Ocean and those flowing westwards via the Yass, Lachlan and other river systems into the Murray Darling Basin. A mostly abrupt escarpment (the Great Escarpment of Ollier 1982) separates the Tablelands from the narrow coastal plain along the present-day east coast. Climates above ~600 m elevation on the Southern Tablelands are cool temperate (microtherm) but relatively dry owing to elevation and rain-shadow effects. The annual precipitation of 500–900 mm supports open woodlands dominated by *Eucalyptus* spp. above an understorey of grasses (Poaceae), daisies (*Asteraceae*) and sclerophyll shrubs such as blackthorn (*Bursaria spinosa*), native cherry (*Exocarpos*) and wattles (*Acacia* spp.). This vegetation type (Southern Tableland Grassy Woodland of Keith 2004) merges into: (i) taller *Eucalyptus* dry sclerophyll forests on stony or infertile soils, *Eucalyptus* wet sclerophyll forest and warm temperate rainforest in wetter areas along the edge of and below the escarpment; and (ii) *Eucalyptus* subalpine woodland from ~1000 m elevation up to the physiological limit of trees at 1800 m. Riverbanks were lined by river-oaks (*Casuarina cunninghamii*) prior to European clearing of the upland plains for grazing in the early–mid 19th century (Burbidge & Gray 1976).

Uplift of the eastern highlands in general, and the Southern Tablelands in particular, remains a matter of scientific controversy, with debate centred on the tectonic relationship(s) of the inland plateaux and

Figure 2 West to east geological cross-section across Lake George basin (from Abell 1985, 1991).
adjoining coastal plains and the chronological implications of relict, commonly deeply weathered land surfaces found at all elevations. For example, unresolved issues highlighted in Ollier & Pain (1994), Brown (2000a, b), Young (2000), Holdgate et al. (2011), Finlayson (2008) and Vandenberg (2010, 2011) include: (a) Are differences in elevation purely owing to erosion of tectonically stable highlands that had already been uplifted in the late Paleozoic to Mesozoic? (b) Were the coastal plain and continental shelf down-warped relative to the highlands following rifting and separation of the Lord Howe Rise from the eastern margin of Australia during the Cretaceous break-up of Eastern Gondwana and the Tasman Sea opening (65–52 Ma)? Mechanisms to explain epeirogenic uplift include crustal thickening by the attachment of a light fraction of igneous mantle material to the continental crust under eastern Australia during opening of the Tasman Sea before ca 80 Ma, and basalt-underplating associated with volcanism during the passage of the Australian plate over a mantle plume after ca 80 Ma (references in Johnson 1989). Both hypotheses provide for an increase in topographic relief during the early Paleogene while allowing the individual remnants of a formerly more extensive Cretaceous paleoplain (‘High Plain’) to have different tectonic histories and have undergone different rates of erosion (see Finlayson 2008; Norvick 2011). Cases in point are the lower Blue Mountains, where proposed uplift in the Paleogene is younger than that of the Great Dividing Range further to the west (Fergusson et al. 2011), the Monaro region south of Canberra, where basalts were extruded onto a plateau whose minimum elevation in the late Paleocene was ~400 m asl (Taylor et al. 1990), and the Eastern Highlands of Victoria where kilometre-scale uplift may have occurred in the mid–late Campanian, with elevations reaching up to 900 m asl by the late Eocene (Vandenberg 2010).

In contrast, agreement exists that the present-day broadly domed-shaped east–west profile of the eastern highlands (and Neogene re-activation of dormant Paleozoic basement faults) reflects the establishment of the northwest–southwest orientated compressive stress regime across Southeast Australia at about 10–5 Ma (late Miocene) following a major reorganisation of plate boundary forces acting on eastern Australia (Dickinson et al. 2002; Sandiford et al. 2004; Clark et al. 2014). Whether neotectonic activity at a particular locality such as Lake George has been more or less continuous since the late Miocene, or has been punctuated by prolonged periods of tectonic quiescence, is not clear. Moreover, the impact on local relief will be obscured except where long-term uplift exceeds local erosion rates or where resistant bedrocks have been brought to the surface (Clark et al. 2014).

The geology and geomorphology of the Southern Tablelands, in the Lake George region, have been described by Abell (1985, 1991), who subdivided the landscape into two major geomorphic units whose boundaries are defined by major fault systems: Plains (paleoplain remnants) between 550 and 750 m elevation, characterised by low-lying, undulating terrain with residual hills and ridges; and Uplands (monadnocks) between 750 and 1100 m elevation, characterised by broad north–south aligned ranges rising up to 500 m above the valley floors. Abrupt linear ‘steps’ in the landscape are considered to reflect tectonic activity and episodes of volcanic activity in the Eocene, Oligocene and early Miocene. The Lake George basin has been considered to be a half graben (e.g. Taylor 1907; Abell 1991) although it remains unclear whether the 120–220 m high escarpment of the Lake George Range (Figures 2, 3) that forms the eastern margin of the associated horst (Cullarin Tableland), is due to neotectonic activity (Singh et al. 1981; Abell 1991) or differential erosion (Garretty 1937; Jennings 1954).

At present, the Lake George basin is a hydrologically closed depression with its western margin being the Lake George Range, and its eastern margin being a line of low hills (Rocky Pic Upland), that form the local watershed of the Great Dividing Range (Coventry 1976; Abell 1985). Streams in the catchment occupy broad valleys in their upper reaches and then meander across alluvial plains such as the Bungendore Plain before reaching the lake.

Geary’s Gap, a topographic low in the Lake George Range (Figure 1) at about 35–40 m above the present lake floor, has been suggested to have been part of the fluvial system draining the Rocky Pic Upland prior to Neogene uplift of the range as well as a paleo-spillway, which connected the basin to the Yass River prior to or during one or more late Quaternary lake highstands (Coventry 1976; Abell 1985). There is no direct field evidence to support the latter claim beyond the general correspondence in height between Geary’s Gap and the highest lake shoreline deposits mapped by Coventry (1976). However, we note that the bedrock topography beneath Lake George provides evidence for a moderately dissected paleo-landscape in which the paleodrainage converges towards Gearys Gap (Figure 3). Ferruginous quartz-rich fluvial gravels occur some 30 m higher and less than 1 km to the south of Geary’s Gap (Figure 3). Similar relict fluvial gravels form local ferruginous capping on the hills, decreasing in elevation to the west of Geary’s Gap, and may represent a former tributary of the Yass River prior to movement on the Lake George Fault (D. Clark pers. comm. 2015) that is currently believed to have created the Lake George basin. In the north and south of the basin, extensive colluvial slope-wash deposits mantle the steep eastern slopes of the Lake George Range. In the past, these extended up to 2 km onto the ancient lake floor (Abell 1991, p. 58) but, as a result of lake-shore erosion, in the central western basin they are now reduced to small alluvial fans at the base of the escarpment (Singh et al. 1981).

The basin is infilled with up to ~165 m of sediments overlying Ordovician bedrock (Figures 2–4). These sediments are subdivided into three major units, in order from youngest to oldest: Bungendore Formation, Ondyong Point Formation and Gearys Gap Formation (Abell 1985, 1991). Bungendore Formation (0–54 m in Core C354) consists dominantly of lacustrine clays, whereas Ondyong Point Formation (54–106 m) consists of fluvo-lacustrine sands, silts and clays (Abell 1985, 1991). These facies are underlain by ~58 m of deeply weathered fluvial sands and gravels (Gearys Gap Formation), believed to have begun accumulating in the paleo-channels of tributaries to the present-day Yass River,
prior to uplift of the Lake George Range. The underlying basement rocks are strongly folded quartz-rich turbidites of the upper Ordovician Pittman Formation and upper Silurian volcanoclastics (Abell 1985, 1991; Finlayson 2008).

PREVIOUS CHRONOSTRATIGRAPHIC STUDIES

Lithostratigraphic and chronostratigraphic data from the Lake George basin, including that from BMR Core hole C354, have been discussed by Singh et al. (1981), Truswell (1984, 1985), Abell (1985, 1991), Singh & Geissler (1985) and McEwan Mason (1991). Cenozoic deposits submitted for paleomagnetic, isotopic and palynostratigraphic dating include gravel cappings on the Lake George Range and fluvio-lacustrine sediments. The former were unable to be dated but are suggested to correlate with deeply weathered Neogene sediments intersected in Core hole C354 and other drill holes in the Lake George basin. Silcrete overlying wave-cut bedrock at the base of the escarpment, south of Geary’s Gap, appears to post-date ferruginous weathering of the gravel cappings. In contrast, core samples of unweathered fluvio-lacustrine facies, infilling the Lake George basin, have provided Neogene to late Quaternary
paleomagnetic, radiocarbon and palynostratigraphic ages of varying reliability (Singh et al. 1981; Truswell 1984, 1985; Singh & Geissler 1985; McEwan Mason 1991). Key core holes include:

1. **ANU Core hole LG4**, drilled in the embayment forming the northern end of Lake George (Figure 1). Here, the base of the Quaternary, marked by the base of the Matuyama Reversed Polarity Chron (ca. 2.58 Ma), occurs at 30.6 m depth (Singh et al. 1981). Underlying sediments to the total depth of 36 m are dated as Gauss Normal Polarity Chron (Figure 4). The current age limits of the latter chron (Chron C2n) are ca. 3.60 to 2.58 Ma, corresponding with the late Pliocene Piacenzian Stage (Lourens et al. 2004; Ogg et al. 2008).

The Gauss/Matuyama boundary coincides with the deposition of deep-water clay facies following a period during which slope wash debris was transported onto the dry lake floor. By extrapolating paleomagnetic and lithostratigraphic data from this core hole to the adjacent 72 m deep but undated BMR Core hole C4, Singh et al. (1981, p. 451) have proposed that lacustrine conditions were ‘already present’ in the late Miocene and therefore the formation of the Lake George basin ‘must be Middle Miocene or older:’

2. **BMR Core hole C354**, drilled ~8 km due south of Core hole LG4 in Lake George and ~2 km east of the escarpment (Figure 1). Magnetostratigraphic analysis confirmed sediments had been deposited during the Brunhes, Matuyama and Gauss Chrons, with the Gauss-Matuyama Chron boundary occurring at 90 m depth (McEwan Mason 1991; see Figure 4). Microflora recovered from core samples of Gearys Gap Formation by Truswell (1984, 1985) yielded a provisional late Miocene age (Assemblage B at 107/C0 110.25 m depth), based on occurrences of *Monoctoca* (fossil equivalent *Monotocidites galeatus*) and *Symplocos* (*Symplocoipollenites austellus*), respectively. Both morphospecies are now known to have extended age ranges in the Gippsland and Murray basins in southeast Australia (Macphail 1999; Partridge 1999, 2006).

**LITHOSTRATIGRAPHY**

The Gearys Gap Formation lacks a fully cored type section and its lithostratigraphy is based on representative sections between 116.5–164.5 m in BMR Core hole C354 where core recovery is relatively good (Abell 1991, pp. 57–58). The unit is suggested to predate the formation of the Lake George basin and therefore its subsurface distribution (Figure 2) may be linked to the reorganisation of the Yass River paleodrainage system.
(Abell 1991, p. 58). While this may apply to the initial phases of accumulation in a fluvial environment, the ~58 m thickness of the Gearys Gap Formation in Core hole C354 is evidence that deposition continued within a basinal setting. The formation comprises deeply weathered, horizontally bedded sands and subrounded, poorly sorted quartz-rich gravels and minor silt and clay strata (see Abell 1991, figure 18), characterised by high levels of kaolinite derived from pervasive weathering of feldspars (Abell 1985, 1991). Ferruginous weathering also is common (Figure 4). The lower boundary of the formation is an unconformity on weathered Ordovician metasediments. According to Abell (1991, p. 58), the upper boundary is ‘arbitrarily taken at the top of the deep weathering profile’ (Figure 4), which Abell et al. (1985) placed at a depth of 106 m in core C354. This boundary corresponds to a change from residual to transported kaolinite, which was interpreted by Abell (1985, 1991) to represent a possible disconformity at the base of the overlying fluviolacustrine Ondyong Point Formation. The formation may correlate, in part, with the patches of ferruginised quartz gravels on the escarpment near Gearys Gap. Silicified gravels (silcrete), which outcrop on the western shoreline of modern Lake George (Figure 3), have developed across a wave-cut bench on the uplifted Lake George Range and therefore are younger than the ferruginous gravels.

METHODS

Three lines of independent age control are used to date particular facies within Gearys Gap Formation and overlying Ondyong Point Formation in BMR Core hole C354. These are: (1) cosmogenic nuclide burial age techniques, to date sediments at 183.75–164.6 m depth and 89.3–90.45 m depth (this study), (2) published palaeomagnetic data (McEwen Mason 1991), and (3) new palynostratigraphic data from carbonaceous sediments at 106–110 m depth (this study). We note that palynostratigraphy and palynostratigraphy have been widely used to provide age control for sedimentary sequences across Australia for the past 3–4 decades. Hence, their application here needs no justification here beyond emphasis in the following: (i) McEwen Mason (1991) was able to recognise a number of magnetic reversals relating to the Brunhes, Matuyama and Gauss Chrons in Core C354. (ii) Biozones defined by the first and last appearance of fossil pollen, spores and other plant microfossils have been shown to provide a reliable method for identifying, dating and correlating Neogene terrestrial deposits in southern Australia, including in the major continental margin and epicontinental basins (see Macphail 1999, 2007; Partridge 1999, 2006). (iii) Palynostratigraphic evidence underpins much of what is currently known about Cenozoic climates (especially precipitation), flora and vegetation across southern Australia (references in Hill 1994; Macphail 1997, 2007).

In contrast to the use of cosmogenic nuclides to date pre-Quaternary deposits in Australia is novel. The burial dating technique (see Granger & Muzikar 2001) is centred on the continuous bombardment of rocks exposed at the Earth’s surface by energetic secondary cosmic rays, principally neutrons and muons. These energetic particles induce nuclear reactions with the constituents of the rock, creating inter alia long-lived radioactive isotopes. If the rock contains quartz (SiO₂), then aluminium-26 (²⁶Al, half-life 720 ka) and beryllium-10 (⁰⁰Be, half-life 1.39 Ma) are produced within the quartz crystal structure from silicon and oxygen, respectively. Rocks exposed at the surface will acquire equilibrium concentrations of these two isotopes that are determined by the balance between production and loss owing to both erosion and decay, i.e. an erosion rate can be deduced from a measurement of the concentration of either radioisotope. If the rock were subsequently removed by fluvial processes, deposited elsewhere and buried at depths where cosmic ray induced production effectively ceases, then the ²⁶Al and ⁰⁰Be would begin to decay. Because they have different half-lives, ²⁶Al decays more rapidly than ⁰⁰Be, and over time the ²⁶Al/⁰⁰Be declines from its original value of approximately 6 (the ratio of production rates of the two isotopes in quartz). The ²⁶Al/⁰⁰Be ratio is therefore a sensitive chronometer for the time of burial. Once the burial age has been determined from this ratio, the concentration of ²⁶Al (or ⁰⁰Be) can be extrapolated back to the time of burial, which allows a determination of the erosion rate of the surface from which the buried material was originally derived. As far as we are aware, this study is the first to apply cosmogenic nuclides to date deeply buried quartz gravel deposits in Australia.

Cosmogenic nuclide samples

Subangular to subrounded quartz pebbles up to 4 cm in maximum diameter were sampled at 163.75–164.6 m depth (Cores 159–160) at the base of the Gearys Gap Formation, and at 89.3–90.45 m depth (Cores 100–101) ~9 m above the boundary between the Ondyong Point and Gearys Gap formations at 106 m depth. These pebbles were crushed and sieved, and the 250–500 μm fraction was taken for analysis. The quartz was purified, followed by the extraction of the ²⁶Al and ⁰⁰Be, in the cosmogenic isotope laboratories at the University of Washington (UW) using the methods described on their website http://depts.washington.edu/cosmolab/chem.shtml. Approximately 0.25 mg of beryllium carrier was added to each sample. The purified quartz contained sufficient aluminium so that carrier was not required. In this case, however, it is necessary to determine the Al concentration, and this was performed by ICP-OES at UW. Measurements of the ⁰⁰Be/Be and ²⁶Al/Al ratios were performed by accelerator mass spectrometry using the 14UD pelletron accelerator of the Heavy Ion Accelerator Facility at the Australian National University and methodologies described by Fifield et al. (2007, 2010). From these ratios, and the known amounts of beryllium (as carrier) and aluminium (from ICP-OES) in each sample, the ⁰⁰Be and ²⁶Al concentration (atoms/g) in the original quartz could be derived, and the ²⁶Al/⁰⁰Be ratio deduced. Burial ages were calculated from the formulae of Granger & Muzikar (2001), using the reference production rate of ⁰⁰Be to be 4.96 atoms/g (quartz)/year at sea-level and high-latitude, a ²⁶Al/⁰⁰Be production ratio of 6.1, and an effective attenuation length for production by
high-energy spallation in rock of 160 g/cm² (see Balco et al. 2008). The altitude and latitude-scaling scheme of Stone (2000) was employed, and a rock density of 2.6 g/cm³ was assumed.

Paleomagnetic samples

McEwan Mason’s (1991) magnetostratigraphic analysis of BMR Core C354 was based on 210 samples taken at intervals less than 1 m spacing along the core, except for six intervals of less than 2 m each between 64.9 m and 84.7 m where core was not recovered or found to be missing. For all measurements, orientation was related to the top and base of each core section, and in the absence of other orientation information only the magnetic inclination was measured. Magnetic remanence directions were measured on a two-axis ScT cryogenic magnetometer at the ANU Black Mountain paleomagnetic laboratory, with samples undergoing stepwise alternating field demagnetisation to isolate their Characteristic Remanent Magnetisations (McEwan Mason 1991). The magnetic stability of each sample was assessed using demagnetisation curves, orthogonal plots and stereoplots (Zijderveld 1967).

**Figure 5** Age-diagnostic morphospecies (family/genus of nearest living equivalent taxon in parentheses). (a) *Cyatheacidites annulatus* (Lophosoriaceae: *Lophosoria*); (b) *Densoisporites implexus* (Selaginellaceae: *Selaginella uliginosa*); (c) *Acaciapollenites octosporites* (Mimosaceae: *Acacia*); (d) *Nothofagidites falcatus* (Nothofagaceae: *Nothofagus* (Brassospora) sp.); (e) *Thymelaeapollis* sp. (Thymelaceae: *Pimelea*); (f) *Sympliocipollenites austellus* (Sympliocaceae: *Symplocos*); (g) *Tricolpites* sp. (Brassicaceae); (h) *Tubulifloridites pleistocenicus* (Asteraceae: *Cassinia arcuata*); and (i) *Tubulifloridites antipoda* (Asteraceae: *Tubuliflorae*).
Palynostratigraphic samples

Plant microfossils preserved in five core chips of unweathered carbonaceous sandy clay (Core 116) and underlying dark-grey laminated claystone (Core 117) between 108 m and 110 m depth, the same interval analysed by Truswell (1984, 1985), were extracted using standard chemical and micro-sieving techniques (see Traverse 1988) by Core Laboratories (Australia) Pty. Ltd in Perth. The maximum and minimum age limits of these samples were determined using the time-distributions of short-ranging fossil pollen and spore species (morphospecies) in the offshore Gippsland Basin in Bass Strait, and in the epicontinental Murray Basin in south-west NSW and Victoria (Macphail 1999; Partridge 1999, 2006). Reconstructions of the depositional environments, vegetation and climates in the vicinity of Lake George are based on the stratigraphic distribution and relative abundance of all formally described fossil algal cysts and miospores recovered from the unweathered carbonaceous section between 108 and 110 m depth. Estimates of relative abundance, calculated as a percentage of the total identifiable miospore count excluding algal cysts

Figure 6 Paleoenvironmental indicators (family/genus of nearest living equivalent taxon in parentheses). (a) Circulisporis parvus (Zygnemataceae); (b) Debarya sp. (Zygnemataceae); (c) Stereisporites australis (Sphagnaceae: Sphagnum); (d) Gleicheniddites sp. (Gleicheniaceae); (e) Cyathidites australis (Cyatheaceae: Cyathea); (f) Trilites tuberculiformis (Dicksoniaceae); (g) Cyperaceaepolis neogenicus (Cyperaceae); (h) Droseridites sp. (Droseraceae: Drosera); and (i) Milfordia hypolaenoides (Restionaceae).
and hornwort spores, are given in Table 2. Rare taxa (relative abundance <1%) and uncertain identifications are indicated by ‘x’ and ‘cf.,’ respectively. Nearest living relatives (NLRs) of the fossil morphospecies are given in the same table. Specimens used to infer the geological age, depositional environment and dryland climate of age diagnostic and environmentally significant morphospecies are illustrated in Figures 5 and 6, and Figures 7 and 8, respectively.

RESULTS

Age control

COSMOGENIC NUCLIDE (10Be/26Al) BURIAL AGE LIMITS

10Be and 26Al concentration data and burial ages and erosion rates inferred from the ratio of the cosmogenic nuclides are given in Table 1. Burial ages of 3.93 ± 0.36
Figure 8 Sclerophyll morphospecies (family/genus of nearest living equivalent taxon in parentheses) (a) *Acaciapollenites myriosporites* (Mimosaceae: *Acacia*); (b) *Banksieaeidites* sp. (Proteaceae: *Banksia*); (c) *Haloragacidites harrisi* (Casuarinaceae); (d) *Gothanipollis* sp. (Loranthaceae); (e) *Malvacipollis spinyspora* (Euphorbiaceae: *Micrantheum*); (f) *Myrtaceidites eucalyptoides* (Myrtaceae: *Eucalyptus*); (g) *Palaeocoprosmadites zelandiae* (Rubiaceae: *Coprosma*); (h) *Paripollis* sp. (Epacridaceae: *Epacris*); (i) *Hakeidites* sp. (Proteaceae: *Grevillea/Hakea*); (j) *Proteacidites* sp. cf. *P. adenanthoides* (Proteaceae); (k) *Proteacidites punctiporus* (Proteaceae: *Stirlingia*); and (l) *Stephanocolpites oblatus* (Haloragodendron/Glischrocaryon).
Table 1  Cosmogenic nuclide ages

| Core no. | Sample data | 10Be data | 26Al data |
|----------|-------------|------------|------------|
|          | Quartz sample mass (g) | ²⁹Be carrier mass (µg) | ¹⁰Be/²⁹Be (× 10^-15) | ¹⁰Be (10^3 atoms/g) | ²⁶Al mass (g) | ²⁶Al/²⁷Al (× 10^-15) | ²⁶Al (10^3 atoms/g) |
| 100–101  | 32.100 | 252.1 | 262 ± 13 | 137 ± 7 | 1.29 | 210 ± 15 | 185 ± 14 |
| 159–160  | 30.042 | 251.1 | 78 ± 5 | 43.0 ± 2.8 | 1.32 | 41.9 ± 4.1 | 37.3 ± 4.5 |

| Core no. | Depth (m) | ²⁶Al/¹⁰Be Burial age (Ma) Erosion Rate (m/Ma) |
|----------|-----------|-----------------------------------|
| 100–101  | 89.3–90.45 | 1.36 ± 0.12 | 2.97 ± 0.23 | 4.75 ± 0.83 |
| 159–160  | 163.75–164.64 | 0.87 ± 0.12 | 3.93 ± 0.36 | 9.7 ± 2.6 |

Table 2  Relative abundance of fossil plant microfossil taxa in BMR Corehole C254 Values are expressed as a percentage of the total identifiable fossil pollen and spore count; those given in parentheses are based on statistically unreliable pollen counts. 'x' indicates relative abundance values less than 1%. 'cf.' indicates morphotypes whose identification is uncertain.

Taxa included in the pollen sum

| Fossil taxon | Nearest living relative | Core number | 116 | 117 |
|--------------|-------------------------|-------------|-----|-----|
|              |                         | 108.1 m     | 109.1–109.2 m | 109.4 m | 110.0 m | 110.3 m |
| Gymnosperms  |                         |             |     |     |
| Araucariacites australis | Araucaria | x | x | x | x | x |
| Dacrycarpiites australiensis | Dacrycarpus | x | x | x | x |
| Dacrydiunites florinii | Dacrydium | x | 2% | x | x | x |
| Dilwynites granulatus | Agathis/Wollemia | cf. |
| Microcarphydites antarcticus | Microcarphry tetragona | x |
| Microalatidites palaeogenicus | Phyllocladus | x | x | x |
| Phyllocladidites mausingii | Lagarostrobus franklinii | x |
| Podocarpidites spp. | Podocarpus–Prumnopitys | 8% | 12% | 11% | (7%) | x |
| Podosporites microsaccatus complex | Extinct clade | x | x | x |
| Podosporites erugatus | Pherosphaera/Microstrobos | cf. |
| Total | 8% | 14% | 11% | (7%) | x |

Angiosperms (woody)

| Fossil taxon | Nearest living relative | Core number | 116 | 117 |
|--------------|-------------------------|-------------|-----|-----|
| Acaciapollenites myriosporites | Acacia | x |
| Acaciapollenites octosporites | Acacia baueri-type | x |
| Banksieaeidites arcaustas | Musgravea | x | x |
| Banksieaeidites elongatus | Banksia | x |
| Eripiptes spp. | Ericaceae | x | x | (1%) | x |
| Gyropollis psilatus | Gyrostemonaceae | x |
| Haluraegidites harrisi | Allocasuarina/Casuarina | 11% | 23% | 20% | (14%) | (2%) |
| Malvaccipollis spinyspora | Micranthemum | 5% | 2% | 8% | x |
| Myrtaceidites eucalyptoides | Eucalyptus (sensu lato) | x | x | x | (3%) |
| Myrtaceidites leptospermoidees | Leptospermum-type | x |
| Myrtaceidites protrudiporens | Myrtaceae | x | (1%) |
| Nothofagidites asperus | Nothofagus (Lophozonia) | 7% | 5% | 3% | (11%) |
| Nothofagidites brachyspinulosus | Nothofagus (Fusospora) | x | 2% | x | (1%) |
| Nothofagidites emarcidus-complex | Nothofagus (Brassospora) | 1% | 9% | 7% |
| Nothofagidites falcatus | Nothofagus (Brassospora) | x | 3% | 1% | (2%) |
| Nothofagidites cf. vansteenisii | Nothofagus (Brassospora) | x | 2% |
| Palaeocoprosmadites zelandiae | Coprosma-type | 1% |
| Paripollis sp. | Epacris sp. | x | x | x |
| Proteacidites obscurus | c. Agastachys | c. |
| Proteacidites persoonioides | Persoonia | x |
| Proteacidites truncatus | Isopogon | x |

(continued)
Table 2 (Continued)

| Fossil taxon                          | Nearest living relative | Core number | 108.1 m | 109.1–109.2 m | 109.4 m | 110.0 m | 110.3 m |
|--------------------------------------|-------------------------|-------------|---------|---------------|---------|---------|---------|
| Proteacidites spp.                   | Proteaceae              | x           | 1%      | x             | x       | x       | x       |
| Pseudowinterapolis spp.              | Winteraceae             | x           |         |               |         |         |         |
| Rhoipites ampereaeformis             | *Amperea xiphocladota*   |             |         |               |         |         |         |
| Rhoipites muehlenbeckiaformis        | *Muhielenbeckia*         |             |         |               |         |         |         |
| Rhoipites (Tricolporites) cooksonii  | Unidentified            |             |         |               |         |         |         |
| Stephanocarpites oblatus             | *Haloragodendron*-type  | x           | x       | x             |         |         |         |
| Symplacopollenites austellus         | Symlocos                | 1%          |         |               | x       |         |         |
| Thymelaeasporus sp.                  | *Pimelea*               |             |         |               |         |         |         |
| Tricolporites (Dodonaeae) sp.        | *Dodonaea viscous-type*  |             |         |               |         |         |         |
| Tricolporopollenites esobaltes       | *Macaranga/Mallotus*     |             |         |               |         |         |         |
| Tubulifloridites antipoda            | Asteraceae (Tubuliflorae)| x           |         |               |         |         |         |
| Tubulifloridites pleistocenicus      | *Cassinia arcuata*-type  |             |         |               |         |         |         |
| **Total**                            |                         |             | 26%     | 45%           | 41%     | (33%)   | (2%)    |

**Angiosperms (herbaceous)**

| Chenopodiapollis chenopodiaceoides   | Chenopodiaceae           | x           |         |               |         |         |         |
| Cyperaceapollis spp.                 | *Cyperaceae*             | 10%         | 3%      | 2%            | (3%)    |         |         |
| Droseridites sp.                     | *Drosera*                |             |         |               |         |         |         |
| Graminidites spp.                    | Poaceae                  | x           |         |               |         |         |         |
| Milfordia hypolanoides               | Restionaceae             | 8%          | 15%     | 8%            | (23%)   | (2%)    |         |
| Tricolporites (Brassicaceae) sp.     |                         |             |         |               |         |         |         |
| **Total**                            |                         |             | 18%     | 18%           | 10%     | (28%)   | (2%)    |

**Angiosperms (unassigned)**

| Rhoipites spp.                       | Unknown                  |             |         |               |         |         |         |
| Tricolporites spp.                   | Unknown                  |             |         |               |         |         |         |
| Tricolporopollenites spp.            | Unknown                  |             |         |               |         |         |         |
| **Total**                            |                         |             | x       | x             | x       | x       | x       |

**Fern and fern allies**

| Bacculatisporites disconformis       | Hymenophyllaceae         |             |         |               |         |         |         |
| Baccalatissporites spp.              | Anthoceratae             | x           |         |               |         |         |         |
| Cyathacidites annulatus              | Lophosoria               | x           |         |               |         |         |         |
| Cyathidites australis/minor          | Cyatheatceae/Lindsaeaceae| x           | 1%      | x             |         |         |         |
| Cyathidites paleospora               | Cyathea                  | x           | x       | x             |         |         |         |
| Dictyophyllidites cf. arcuatus       | Gleicheniaceae           | 4%          | x       | x             |         |         |         |
| Poveotrilites balteus                | Lycopodium               |             |         |               |         |         |         |
| Gleicheniidites spp.                 | Gleicheniaceae           | 38%         | 20%     | 31%           | (26%)   | (92%)   |         |
| Laeogatosporites ovatus/major        | Blechnaceae/Scizaeaceae  | x           | x       | x             |         |         |         |
| Latoobosporites marginis             | *Lycopisella lateralis*  | x           | x       | x             | x       |         |         |
| Neorastrikkia equals                 | Hepaticae                |             |         |               |         |         |         |
| Retitrilites austrorovelatidites      | Lycopodiaceae            | x           | x       | x             |         |         |         |
| Trilites tuberculiformis             | Extinct *Dicksonia* sp.  | x           | x       | x             |         |         |         |
| Unassigned trilete spores            | Unidentified             | x           | x       | 1%            | x       | x       |         |
| **Total**                            |                         |             | 42%     | 21%           | 32%     | (26%)   | (92%)   |

**Pollen sum**

|                    |                | 302 | 320 | 258 | 99 | 100 |

(continued)
Ma and 2.97 ± 0.23 Ma for quartz gravels at 163.75–164.6 m depth and 89.3–90.45 m depth, respectively, provide a maximum age limit of early Pliocene (Zanclean) and a minimum age limit of mid late Pliocene (Piacenzian) for the deposition of Gearys Gap Formation at the core hole site.

PALEOMAGNETIC AGE LIMITS

Magnetostratigraphic data date the sampled intervals at 108–110 m depth to the Gauss Normal Polarity Chron (see Figure 4). Current age limits for the upper part of this Chron, above the Kaena Subchron, are 3.03 Ma and 2.58 Ma. This period equates to the late Pliocene, following the 2009 definition of the start-date of the Pleistocene from 2.59 Ma to 1.61 Ma. The new boundary definition makes the former late Pliocene Gelasian Stage (2.59–1.81 Ma) the first Stage of the Pleistocene Epoch (and the Quaternary Period), and the former ‘mid’ Pliocene Piacenzian Stage (3.60–2.58 Ma) the last Stage of the Pliocene Epoch (see Ogg et al. 2008). Sediments below a depth of ~115 m yielded inconclusive results, owing to erratic paleomagnetism, but extrapolation of sedimentation rates yields an age estimate of ca 4 Ma for the base of the core (Figure 4), consistent with the cosmogenic nuclide burial ages.

PALYNOSTRATIGRAPHIC AGE LIMITS

Differences in lithology between the sandy clay unit sampled at 108.1–109.4 m depth (Core 116) and the laminated claystone sampled at 110.0–110.3 m (Core 117) might imply a hiatus in sediment accumulation but any time gap is considered to be minimal given all assemblages are dominated by the same small group of fossil pollen and spores (Table 2).

Maximum age limit: The maximum age of the carbonaceous sands and claystone between 108.1–109.4 m in depth in the Gearys Gap Formation is inferred to be late Pliocene (ca Piacenzian) based on two species (NLRs in parentheses) that first occur in the late Pliocene—early Pleistocene Tubulifloridites pleistocenicus Zone in the Murray Basin, viz. Thymelaeopsis sp. (Pimelea) at 108.1 m depth and Accaciapollenites octosporites (Acacia sp.) at 109.4 m depth. The same maximum age limit is supported by a reticulate tricolpate morphotype resembling Brassicaceae pollen at 109.1 m. If age-range data for the Gippsland Basin (Partridge 1999) were used, then the maximum age could be as old as early Pliocene, based on the first occurrence of Tubulifloridites pleistocenicus (Cassinia arcurata-type) in the Myrtaceidites lipsis Zone (see Partridge 2006) or as young as earliest Pleistocene based on the first appearance of Accaciapollenites octosporites in the overlying Tubulifloridites pleistocenicus Zone in that basin. We note that the microflora include a number of other taxa that first occur in the early to mid Pliocene (Zanclean—Piacenzian) Myrtaceidites lipsis Zone in the Gippsland Basin and its correlative, the Monotocidites galeatus Zone, in the Murray Basin. Examples are Densoisporites implexus (Selaginella uliginosa), Dodonaea-viscosa-type and Rhoipites ampereiformis (Ampherea xiphoclada).

Minimum age limit: The minimum age of the carbonaceous sand and claystone units is unlikely to be younger than Pliocene, based on Cyatheacidites annulatus (Lophosoria), a widely dispersed spore that last occurs in the Myrtaceidites lipsis Zone in the Gippsland Basin and its correlative, the Monotocidites galeatus Zone in the Murray Basin. The same minimum age limit is supported by significant relative abundances of Nothofagidites falcatus between 109.1–110.0 m depth, since this fossil morphospecies (NLR Nothofagus subgenus Brassocpora) became regionally extinct across much of southern Australia by the Plio-Pleistocene boundary. A Pliocene (mid Piacenzian) minimum age is confirmed by the cosmogenic burial age of ca 2.96 Ma ± 0.23 Ma for quartzite gravels at...
89.3–90.45 m depth (see above). Overlying (mottled clays, sands, gravels) and underlying (ferruginous silts, clays) facies are palynologically barren.

**Late Pliocene depositional environments, vegetation and climate**

Three samples (108.1 m, 109.1–109.2, 109.4 m) yielded >250 identifiable specimens and estimates of relative abundance (Table 2) for these are considered to be statistically robust. The other two samples (110.0 m, 110.3 m) yielded fewer than 100 identifiable specimens, and the relative abundance values (given in parentheses) are considered reliable only for commonly occurring taxa such as *Gleichenitidites*.

**DEPOSITIONAL ENVIRONMENT**

Depositional environments can be inferred from the ecology and relative abundance of algal cysts, fern spores and herbaceous pollen recovered from the carbonaceous claystone unit at 108–110 m depth (Figure 6). All samples yielded frequent to common sedge (Cyperaceae) and cord-rush (Centrolepidaceae-Restionaceae) pollen as well as much larger numbers of mature and immature fossil spores (*Gleichenitidites*) of the coral-fern family Gleicheniaceae. The former (herbs) typically dominate mires and shallow swamps, while the latter (ferns) occur in a variety of shaded and open sites but always in damp to wet habitats such as damp creek banks, dune-souls and cliff-top swamps. Despite the parent plants being wind-pollinated, the miospores are poorly dispersed into the surrounding landscape except occasionally by water. Other, but less common, wetland plants included peat-moss (*Sphagnum*), club-mosses (*Lycopodium laterale*), filmy ground-ferns (e.g. *Hymenophyllaceae*) and a sun-dew (*Drosera*). By comparison, the relative abundances of freshwater algal cysts, in particular *Botryococcus* (trace to 2%), are too low to indicate lacustrine conditions and are considered to represent freshwater pools within the fen or equivalent niches along the inflowing streams.

**PALEOVEGETATION**

The dryland pollen influx has been extensively diluted by locally produced spores and pollen. Nonetheless *Nothofagidites* spp. occur in significant numbers and are strong evidence that a now-extinct form of *Nothofagus* temperate rainforest, which included gymnosperms and *Symlocos*, was growing around the Lake George basin into late Pliocene time (Figure 7). Possible niches include river banks (gallery rainforest) or, if the carbonaceous interval between 108 m and 110 m depth post-dates neotectonic activity in the area, shaded gullies on the eastern escarpment of the rising Lake George Range. Other woody pollen types confirm that the dryland vegetation also included dry sclerophyll communities (Figure 8). For example, the most commonly recorded taxon in the microflora, *Haloragacidites harrisii*, almost certainly represents dry sclerophyll species of *Allocasuarina* and/or *Casuarina* and implies that sclerophyll woodland was established on drier hill slopes or that tree or shrub populations of one or both of these genera lined the banks of inflowing streams. Uncommon to rare shrub and/or tree taxa (fossil species in parentheses) such as *Acacia* (*Acaciapollenites myriosporites*, *A. octosporites*), *Banksia* spp. (*Banksieaides cf. B. arcaatus, B. elongatus*), *Dodonaea viscosa*-type (*Dodonaea sphaerica*), *Epicarpidaeae*/*Eriaceae* (*Ericites, Paripollis* ssp.), *Loranthaceae* (*Gothanipollis* sp.) and *Miconia* (*Malaepipollis spinyspora*) may have been part of this woodland or formed separate communities elsewhere on rocky outcrops. Some woody plants appear to have colonised drier sites within the fen. An example is the undescribed gynmae-verberrate morphotype (*Paripollis* ssp.) whose most likely NLR is *Epacris*, a low heath whose pollen are seldom transported far from the parent plant and which still includes mire species over a wide range of elevations, e.g. *E. paludosa* (see Burbidge & Gray 1976). The combined data point to a mosaic vegetation of drought-tolerant sclerophyll and drought-intolerant rainforest plants whose distributions across the basin and adjacent uplands (as now) reflected gradients in soil moisture, aspect and, for fire-sensitive plants, also the local fire frequency.

**WILDFIRE**

Charcoal evidence for paleo-wildfires is widespread globally in terrestrial deposits throughout much of the Phanerozoic (see Bond 2014), and in Australia wildfires are usually implicated in the expansion of flammable sclerophyll biomes during the Neogene (Kershaw et al. 2002; Sniderman & Haberle 2012). In BMR Core CS54, plant detritus (palyndebbris) recovered from the carbonaceous sand and claystone interval is dominated by a mixture of well-preserved but finely disseminated tissues with well-preserved cell walls and very dark brown particles with all trace of the cellular structure lost (semi-opaques). Whether the latter particles are carbonised (micro-charcoal) owing to wildfires or the result of natural oxidative processes is not certain for two reasons. (1) The thin semi-translucent edges of these particles are a mid-brown colour, not grey as would be expected if they were due to pyrolysis. (2) Carbonised xylem particles (inertinite) are extremely rare or absent in most samples. Without independent evidence for burning, e.g. the presence of biomarkers such as levoglucosan (see Dos Santos et al. 2013) all that can be inferred with certainty is that wildfires were present during the Piacenzian but neither the intensity nor frequency was sufficient to eliminate populations of rainforest taxa such as Araucariaceae, Podocarpaceae and Nothofagaceae, most species of which are intolerant of repeated fires (references in Enright & Hill 1995; Veblen et al. 1996), during the period of record.

**PALEOClimate**

Modern climates on the Southern Tablelands of NSW are classified as temperate with uniformly distributed rainfall and prolonged but mild summers (Gentilli 1972; Bureau of Meteorology 1998). At present, neither the annual mean (~625 mm) nor summer (~168 mm) rainfall is adequate to support wet sclerophyll or rainforest communities except on the edge of the Great Escarpment or in sheltered damp gullies in the adjacent ranges. However, modern bioclimatic data for *Nothofagus*, the
dominant tree genus in living cool temperate and montane rainforest, provide a broad-brush guide to conditions prevailing in the Lake George basin during the late Pliocene (cf. Read & Brown 1996; Read & Hope 2005). Precipitation: Distribution data for living Nothofagus subgenus Lophozonia spp. in eastern Australia imply that the mean annual rainfall in the vicinity of Lake George during the Piacenzian exceeded $\sim 1880$ mm, of which a minimum of $350$ mm was received during the summer quarter. Equivalent parameters for living Nothofagus subgenus Brassocarpetes spp. in New Guinea and New Caledonia are $\sim 2890$ mm pa and $\sim 890$ mm pa, respectively. These data imply an annual rainfall during the Piacenzian Stage of $\sim 2000–3000$ mm, although this probably was not uniformly distributed throughout the year.

Temperature: Mean air temperatures are less easily inferred since the annual mean temperature for areas able to support Nothofagus (Lophozonia) spp. in eastern Australia is $9.3^\circ$C (microtherm climate) whereas the equivalent for Nothofagus (Brassocarpites) spp. in New Guinea and New Caledonia is $16.3^\circ$C (mesotherm climate). Which (if either) of these two temperature values is the more applicable to the Lake George area during the Piacenzian is debatable since the altitudinal distribution of both subgenera extends from the coast into the upper subalpine/alpine zones, as do the altitudinal distributions of the gymnosperm genera found in the overstorey in modern temperate rainforest, e.g. Dacrycarpites australiensis (Dacrycarpus), Dacrydiomites florinii (Dacrydium), Microalatidites palaegonicus (Phylocladus) and Podocarpites (Podocarpus–Prumnopitys). Although apparently rare, the presence of Symwoola Flats australinii is, ecologically, in better agreement with mean annual temperatures in the mesotherm range ($>14^\circ$C $<20^\circ$C) during the Piacenzian given that living Symwoola Flats spp. are almost entirely restricted to warm-temperate to subtropical rainforest in Australia, from the South Coast and Central Tablelands of NSW into Queensland. Such relatively warm conditions would explain the anomalously late survival of the Cyathaeidites annulatus and Nothofagidites falcatus at high elevations on the Southern Tablelands.

**DISCUSSION**

The cosmogenic burial age of $3.93 \pm 0.36$ Ma measured for the basal gravels at $\sim 164–165$ m depth is the first for a pre-Quaternary sequence in Australia. The age provides a late Zanclean age for the formation of the Lake George basin, which calls into question (but does not wholly disprove) earlier estimates of late Miocene or earlier for the uplift of the Lake George Range and basin formation. For example, the calculated erosion rates of $\sim 10.3$ m/Myr at $3.93$ Ma and $\sim 4.9$ m/Myr at $2.97$ Ma (Table 2) are consistent with initially high levels of tectonic activity during the late Zanclean (compare Clark et al. 2014). However, we emphasise that the inferred erosion rates assume (i) the subrounded gravels used for cosmogenic dating were at the surface before being buried, i.e. had not been exhumed, and (ii) the exposure was not appreciably shielded by the local topography. Neither caveat can be dismissed, but (i) the orientation and depth (up to $100$ m) of paleovalleys incised into Ordovician bedrock beneath Lake George (see Figure 3), and (ii) rounded–subrounded morphology of some of the quartz clasts suggests that they are most likely to be fluvial, i.e. were deposited by a stream flowing across the study site from the Paleozoic Rocky Pic Upland, not colluvium shed from the Lake George Range escarpment. However, a cosmogenic nuclide burial age of $2.97 \pm 0.23$ Ma (Table 1) for quartz gravels at $89.3–90.45$ m, $15$ m above the base of the overlying Ondyong Point Formation in Core hole C354, confirms that some $75$ m of fluvial and lacustrine sediments had been deposited in the lake basin by mid late Pliocene time. Given that colluvial sediments have been transported up to $2$ km into the lake basin, distinguishing between fluvial and colluvial facies may require shallow seismic techniques to determine the (offlap/onlap) relationships of fluvial gravels within the Gearys Gap Formation and colluvial facies shed from the Lake George Range.

Carwoola Flats, in the headwaters of the Molonglo River, some $20$ km south of Lake George (Figure 1) may represent a modern analogue for the depositional environment of the Gearys Gap Formation. Unlike the paleodrainage at Lake George, the Molonglo River has not been blocked by movements on the Lake George Fault. This has allowed the river to continue to downcut at a sufficient rate to overcome tectonic uplift, thereby forming a steep-sided gorge through the Cullarin Horst. Drilling and seismic refraction surveys (Abell 1991, pp. 62–64) indicate that Carwoola Flats overlies a shallow bedrock depression, up to $30$ m deep, containing a wedge of alluvium and colluvium (including gravels) that thickens westwards, towards the fault. In this case, fault movement has created $30$ m of accommodation space on the eastern (downthrown) side of the fault, but still allowed through-flowing drainage to the west. We envisage that a similar sedimentary model could be applied to much or all of the Gearys Gap Formation.

At present, the earliest confirmed evidence for neotectonic activity on the Southern Tablelands during the Neogene is a K/Ar age of $18.8$ Ma (early Miocene), for basals butting up against the Kurrajong Fault scarp at Green Scrub on the eastern escarpment of the Blue Mountains west of Sydney (Wellman & McDougall 1974). Nothofagidites microflora in alluvial fan deposits that have prograded across the plane of this Fault at Mount Lagoon support the same (Miocene) age (Macphail 2013; McPherson et al. 2014). At Lake George, the combined chronostratigraphic and lithostratigraphic data confirm deposition of the Gearys Gap Formation at the core site began sometime during the Zanclean and ceased sometime during the mid late Piacenzian. Accordingly, we propose that rates of uplift vs fluvial incision on the Lake George Range first became sufficiently high to occlude local drainage lines during the early Pliocene, not Miocene. This conclusion is supported by: (1) the inferred age of the carbonaceous claystone at $109.1–110.3$ m depth, which is only slightly younger than the ‘mid-Pliocene’ age inferred for onset of lacustrine conditions in ANU Corehole LG4 at the northern end of Lake George; and (2) the very low to trace relative abundance of freshwater algae, which is strongly
against a freshwater lake extending across the core hole site. The earliest time a freshwater lake occupied the lake basin remains unresolved although core log data archived at Geoscience Australia show that ~75 m of fluviol and presumed lacustrine sediments had accumulated in the lake basin between the intervals dated (this study) as late Zanclean Stage and mid Piacenzian.

Independently dated Pliocene microflora have not been recorded elsewhere on the Southern Tablelands, but probable correlative is preserved in alluvium infilling valleys on the central-west slopes of NSW and in the adjacent Murray Basin (references in Martin 1987; Macphail 1999). Many of these assemblages resemble microflora recovered from BMR Core hole CS54 in terms of composition although not in pollen dominance (compare Martin 1973). In contrast, the Lake George microflora differ markedly in terms of species composition and/or pollen dominance from older (Paleocene to Miocene) microflora recovered from not-dissimilar mire and lacustrine depositional environments on the Southern Tablelands and surrounding ranges (see Macphail 2007).

CONCLUSIONS

The cosmogenic nuclide burial age for gravels at ~163.75–164.6 m depth and the combined paleomagnetic and palynostratigraphic evidence for lacustrine facies at ~108–110 m depth demonstrate that: (i) alluvium was being deposited at the core site at ca 3.93 Ma (late Zanclean Stage), although it is unclear whether this occurred within a basinal setting; (ii) the Gearys Gap Formation is likely to be a wholly Pliocene formation; and (iii) deposition of the Gearys Gap Formation had ceased in the Piacenzian Stage. Carbonaceous sand and laminated claystone, at ~108–110 m depth, accumulated in a fern wetland (mire), not a freshwater lake, but overlie floodplain deposits. This wetland community might reflect an early stage in the formation of a laterally extensive deepwater lake, analogous to that developed during the Last Glacial Stage of the Pleistocene or merely reflect a lowstand environment in a shallow lake characterised by fluctuating water levels. Conversely, the dryland vegetation on the Lake George Range included the last remnants of the Nothofagus rainforest that had dominated the Southern Tablelands since the early Paleocene ca 53 million years before (see Macphail 2007). The genetic relationship of Lake George with tectonic uplift or another factor such as the progradation of alluvial fans across paleo-spillway(s) linking Lake George to the Yass or Lachlan River systems remains unconfirmed. Nevertheless, the combined chronostratigraphic data imply that the formation of the Lake George basin is more recent than the Miocene or older age proposed by Singh et al. (1981). If the quartz gravel unit at 163.75–164.6 m depth predates uplift of the Lake George Range, then relict quartz gravels on the summit ridge of the Lake George Range could imply neotectonic uplift of 100–200 m in 4 million years, a rate that is not dissimilar to uplift in the most actively deforming part of the Australian continent, such as the Flinders Ranges, and Otway and Gippsland basins (Sandiford 2006a, b; Quigley et al. 2006; Clark et al. 2012). The nature of early and late Pleistocene environments in the Lake George basin will be examined in subsequent papers.

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