Cenozoic stratigraphic succession in southeastern Australia

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Strata of Cenozoic age occur around the southern margin of Australia as thin and discontinuous outcrops, interpolated and fleshed out by economic exploration onshore and offshore. The neritic strata fall into four sequences or allostratigraphic packages of (I) Paleocene – Early Eocene, (II) Middle Eocene – Early Oligocene, (III) Late Oligocene – Middle Miocene, and (IV) Late Miocene – Holocene age: a four-part pattern that can be seen also in the flanking pelagic and terrestrial realms including regolith deep weathering. Problems of correlation and age determination (predominantly biostatigraphic) have included biogeographical constraints (endemism in neritic molluscs and terrestrial palynomorphs, mid-latitude assemblages in calcareous plankton), and slow progress in magnetostratigraphy and chemostratigraphy. Sequence I largely repeats the Cretaceous siliciclastic–coal, marginal-marine facies (carbonate-poor, with marine and non-marine palynomorphs and agglutinated foraminifers) punctuated by marine ingressions with microfaunas and sparse macrofaunas. Sequence II contains the first carbonates in the region since the Palaeozoic and the most extensive coals of the Cenozoic anywhere. Sequence III contains the most extensive neritic carbonates and the last major coals. Sequence IV is more strongly siliciclastic than the two preceding. Each of these four second-order entities (107 years duration) comprises third-order packages each with an unconformity and marine transgression. These packages hold true right along the southern Australian margin in the sense that the hiatuses and transgression do not display significant diachronity at the relevant timescales (105–106 years). Recognised, delimited and correlated independently of the putatively global Exxon sequences, they are remarkably consistent with the latter, thereby providing a significant regional test. There are two widespread emphases on southern Australian geohistory and biohistory: (i) to regard the regional story as part of the global story of accreting continents, an expiring Tethys, and an episodically cooling planet; and (ii) a somewhat contrary emphasis, with the region being a special case of rapid longitudinal motion towards the equator. Both emphases are plausible with the former being more heuristic. The stratigraphic record is strongly punctuated, the four sequences being separated by both tectonic and climatic events. Thus: the sequence I/II gap involved extensive plate-tectonic reorganisation and a new spreading regime from ca 43 Ma, coevally with early growth of Antarctic ice; in the II/III gap, deformation in marginal basins is coeval with a global low in cooling, large ice sheet and falling sea-level to ca 30 Ma; and the III/IV gap is marked by widespread cessation or contraction of stratal accumulation and withdrawal of thermophilic taxa coevally with the major expansion of at 14 Ma of the Antarctic ice sheet, onset of intense canyon cutting, and plate-wide basin inversion.

KEY WORDS: Australia, Cenozoic, climatic change, correlations, marine transgression, palaeo-environments, stratigraphy, unconformities.

INTRODUCTION

Martin Glaessner, the founding Editor of the Geological Society of Australia, had a wide interest in all aspects of geology. He wrote two key papers heralding modern Cenozoic studies in the Australian region (Glaessner 1943, 1951) and supervised numerous theses on Cenozoic stratigraphy and palaeontology in the 1950s and 1960s. Glaessner’s instincts were deeply stratigraphic, and it seems appropriate some half a century later to review the status of our knowledge on the micropalaeontology and stratigraphy of this region in the Geological Society’s journal.

Our own work has focused on the neritic realm as pivotal between the pelagic realm and the terrestrial realm, with its mantle of regolith and the attendant problems of discovering, ordinating and correlating the non-marine floras and faunas, and of relating onshore coals to offshore limestones. Southeastern Australia has several sedimentary basins facing the Southern Ocean (Figure 1). During the Cenozoic the region was not subjected to the levels of deformation and topographic relief found on the northern, active continental margin in New Guinea or across the Tasman Sea in New Zealand. Onshore, strata tend to be thin, almost flat or gently folded and faulted, and outcropping discontinuously, necessitating a piecemeal approach to correlation and ordination. Age determinations and arguments about chronostratigraphic units (especially regional stages and their boundaries) have
never had the urgency that these activities have had in New Zealand, where Cenozoic chronology and correlation down to regional Stage level have been vital to geological mapping. Facies are non-marine to outer neritic, with pronounced thickening and slope deposits in the offshore subsurface. Access to the subsurface onshore has been driven by exploration and development of resources mostly in petroleum exploration, which has moved offshore long
since, and in water, coal, non-metallic minerals, and placers.

The Cenozoic geology of southern Australia has been summarised most recently by Alley and Lindsay (1985), Holdgate and Gallagher (2003) and Holdgate (2006). A persisting theme in regional Cenozoic studies has been the importance of global climatic and palaeoceanographic change (Figure 2), against which the Australian record can be assessed.

CENOZOIC SUCCESSION IN SOUTHERN AUSTRALIA: STRATIGRAPHY

Correlation and age determination: biostratigraphy

As in the classical Tertiary rocks of Europe, the molluscs were prominent in early Australian stratigraphy. It was already known that the modern southern Australian assemblages were highly endemic, as attested by the number of new species described in the famous malacological school in Paris on material from the French voyages to this region at the turn of the 18th/19th centuries; and their Cenozoic antecedents soon proved to be the same. A succession of molluscan assemblages could be assembled (Darragh 1985) and analysed for Lyellian percentages in a rudimentary way, but endemism was a powerful block to correlations with other regions including classical western Europe, as discussed for the Far East long ago by Glaessner (1943). It is apparent from Singleton’s (1941) major review of the Australian Tertiary that fossils did not really play a decisive role in the ordination and correlation of the regional Cenozoic strata as late as the 1930s.

Although foraminifers were being described since the late 19th century, and although Crespin (1943) made some progress in shifting the definition and characterisation of the stages towards foraminifers and lithostratigraphy; it was Glaessner’s (1951) incisive discussion of microfossils and chronostratigraphy that provided the real basis for subsequent progress in correlation and age determination.

Figure 1 (Upper) The geographical configuration at 43–42 Ma (late Middle Eocene). The southern margin with new or rejuvenated basins faced a Gulf or neritic throughway before oceanic breakthrough. (Inset) Indo-Atlantic and Pacific, a biogeographical theory whereby two provinces are merged into the Southern Ocean Province. W-STR, E-STR, west and east components of South Tasman Rise; ETP, East Tasman Plateau. From McGowran et al. (2000) and various sources. (Lower) Australia’s migration on a time–latitude grid, with the continent shown in its Early Cenozoic, mid-Cenozoic and Late Cenozoic latitudes (McGowran et al. 1997a). Superimposed are three oceanic grossplot isotherms (Frakes et al. 1994; Frakes 1999), together showing three warmings in the Paleocene to Early Eocene, late Middle to Late Eocene, and Early to Middle Miocene. These are also times of major episodes of accumulation of coals in southern Australia (arrows). Episodic precipitation (and inferred increased intensity of deep weathering) are predicted by polewards swings in oceanic surface isotherms, not by the trajectory of continental drift through implicitly static climatic belts. From McGowran and Li (1998).

(Carter 1958a,b, 1964; Wade 1964, 1966). Meanwhile, Jenkins (1966) used the most complete Neogene section available in southern Australia, the Lakes Entrance Oil shaft in east Gippsland, to erect an independent planktonic foraminiferal succession, which became a platform for his subsequent major study of the New Zealand Cenozoic succession (Jenkins 1966, 1971, 1985; Hornibrook et al. 1989). The New Zealand planktonic foraminiferal biozonation was found to be applicable, with modifications, in South Australia (Lindsay 1986; Ludbrook & Lindsay 1969; Ludbrook 1971; McGowran 1973a). McGowran et al. (1971) attempted the first comprehensive correlation of regional zones and stages (perhaps anywhere) to a modern, global, numerical geochronology (Berggren 1969). Abele et al. (1976, 1988) reviewed biostratigraphy and chronostratigraphy in southern Australia and their application to the Victorian neritic record, clearly preferring the Carter (1958a, 1964) scheme. Thus, by the 1970s Victorian and South Australian micropalaeontologists were using different zonations based on much the same foraminiferal succession in the neritic realm, while the New Zealand succession comprised yet a third subset. McGowran (1978) presented a correlation of Eocene zones and sub-zones in extratropical southern Australasia with the observation that zones were of decreasing usefulness: that bioevents, initial and final appearances, and temporary immigrations and ephemeral occupations, are what count in correlation.

In petroleum exploration of the western and southern continental margins and especially the Gippsland Basin offshore, instead of a biozonation being tied straightforwardly to electric logs (and becoming more or less redundant in the process), the complexities of multiple canyon cuts and fills demanded a close and ongoing application of biostratigraphy. The ‘letter zones’ developed by Taylor (1983, 1986; the zones were calibrated in McGowran 1978) presented a correlation of Eocene zones and sub-zones in extratropical southern Australasia with the observation that zones were of decreasing usefulness: that bioevents, initial and final appearances, and temporary immigrations and ephemeral occupations, are what count in correlation.

Meanwhile, Stott and Kennett (1990) erected Antarctic Palaeogene (AP) zones; Jenkins (1993) extended his New Zealand schema to cover the Southern Ocean Palaeogene (SP zones) and Neogene (SN zones); and Berggren et al. (1995) defined transitional zones for the Neogene: transitional between the tropical and polar realms as oceanographic gradients steepened in the cooling Southern Ocean. McGowran and Li have tended to follow McGowran’s early decision to use defining events (i.e.
datums), abandoning both stratigraphic ranges and so-called standardised zones, and acknowledging as formally defined entities only the standard, as in the Palaeogene framework here (Figure 3), although Li et al. (2003c, 2004) have defined Neogene zones for the southern Australian succession (SAN zones) (Figure 4).

Calcareaous nannofossils have provided a valuable cross-check on the dating of marine strata including several neritic facies lacking a strong planktonic foraminiferal record (Shafik 1983; Chaproniere et al. 1996). Palynology has been essential in the all-important cross-tying of the marine and non-marine facies (Stover & Evans 1973; Stover & Partridge 1973; Chaproniere et al. 1996; Morgan et al. 1995), but a major study remains unpublished (Partridge 1999). Other methods of correlation and age determination have included radiochronology, magnetostratigraphy and chemostatigraphy, although these methods have had relatively minor roles in the region. This is changing as the results of ODP legs 182 and 189 appear (Holbourne et al. 2004; Li et al. 2004) and as strontium isotopic dating progresses (Dickinson et al. 2001, 2002). A stratigraphic framework for Australia and New Zealand (Figure 5) may be useful for understanding the regional chronostratigraphy and its relationship to classical chronostratigraphy and the third-order sequences.

**Cenozoic biogeography**

Australia has been important to some of the debates about the geography of evolution and distribution: debates such as dispersal of continents vs biotic dispersal, phyletic relationship vs phenotypic convergence, or biotic dispersal itself vs vicariance (McGowran 1973b; Woodward & Case 1996; Crisp et al. 1999). The components of neritic and pelagic biogeography may be summarised as follows.

Australia and New Zealand are at the southern margin of the great Indo-West Pacific bioprovince, which is centred on the Pacific warm pool in the western-tropical part of the Pacific Ocean. There is an entrenched perception that Australia’s drift northwards into warmer climes is balanced by global cooling and an equatorwards retreat

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**Figure 2** Global climatic scenario: four successively weaker warmer times terminated by successively stronger coolings. EECO, Early Eocene Climatic Optimum; MECO, Middle Eocene Climatic Optimum (Bothay & Zachos 2003); LOWE, Late Oligocene Warming Event; CHILL, Late Oligocene Warming Event and ichesheet decay; MICO, Miocene Climatic Optimum. Cooling curve, smoothed δ¹⁸O record of ocean-floor benthic foraminifers (Miller et al. 1987) scaled by Crowley (2000) using tiepoints to give a bottom-water temperature difference from the modern deep ocean. Senonian warm-saline *Inoceramus* ocean, MacLeod et al. (2000); Early Palaeogene recurring hyperthermals?, Thomas et al. (2000); Auversian facies shift, Berger and Wefer (1996) and McGowran and Li (2000); Miocene climatic optimum and Pliocene warm reversal, several recent discussions; Chills I–IV, McGowran and Li (1997a) and Lear et al. (2000). Latitudinal ocean-temperature profiles for surface waters (SST) (lines, scatter omitted) and bottom waters (striped envelopes): modern and Early Eocene (with alternative SST profiles), Bice et al. (2000); Late Maastrichtian and Late Albian–Cenomanian, MacLeod et al. (2000). Modern SST profile repeated for ease of comparison. During global cooling SST profiles steepen and SST/bottom temperature differences increase. Australian scenario, general statements from various sources (Hill 1994; pers. comm. 2002; McGowran et al. 1997a, 2000; Veevers 2000b). Two grey spots with error bars: clustered dates of ferruginisations in the Australian regolith, implying two episodes or intensifications in deep weathering (Pillans 2002).
by oceanic water-masses, tracked by isotherms. Although neither component of environmental change is to be doubted, Figure 1 suggests that much more rapid global climatic shifts of tens of degrees latitude are strongly superimposed on much slower and steadier sea-floor spreading and continental drift.

Southern Australia is extra-tropical (cool-water) and faces the growing Southern Ocean, the engine-room of the cooling planet. Its neritic and pelagic biogeographical archives accumulated as outcomes of balances between Indo-Pacific immigrations during warm times and the advance northwards by the Subtropical Convergence during cooling times (Figure 6) (Gallagher et al. 2003). This record is shared with New Zealand, which has a thicker and more informative stratigraphic succession and a stronger response to the counterclockwise gyre in the southwest Pacific (Hornibrook 1992; McGowran et al. 2000; figure 8). However, southern Australia west of Tasmania has been warmed from the west—against the counterclockwise gyre of the Indian Ocean (uniquely among the southern continents)—by the proto-Leeuwin and Leeuwin Current (McGowran et al. 1997b) (Figure 7).

Li et al. (1996) looked at endemism among the neritic benthic foraminifers in southern Australia and tabulated species in three categories: endemic plus semi-endemic, immigratory, and cosmopolitan. The endemics were mostly shallow-epifaunal, whereas infauna species fall into the cosmopolitan group. The immigratory species all belong in the group of the so-called larger foraminifers: those with photosymbionts, which flourish in warm, shallow, well-lit,

### Figure 3
**Palaeogene foraminiferal–biostratigraphic framework for southern Australia.** Integrated geochronology, Berggren et al. (1995); third-order sequences and dates, Hardenbol et al. (1998); regional transgressions and bioevents, McGowran et al. (1997a; modified); Taylor planktonic foraminiferal zones, Taylor (1983, 1986); pollen zones (Stover & Partridge 1973) as correlated in Chaproniere et al. (1996) and McGowran et al. (1997a). Correlations of three of the four Nothofagidites asperus zonal boundaries are fairly firm but the lowest boundary with the P. asperopolus zone is chronologically very poorly constrained and not shown here.

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**Figure 3**: Palaeogene foraminiferal–biostratigraphic framework for southern Australia. Integrated geochronology, Berggren et al. (1995); third-order sequences and dates, Hardenbol et al. (1998); regional transgressions and bioevents, McGowran et al. (1997a; modified); Taylor planktonic foraminiferal zones, Taylor (1983, 1986); pollen zones (Stover & Partridge 1973) as correlated in Chaproniere et al. (1996) and McGowran et al. (1997a). Correlations of three of the four Nothofagidites asperus zonal boundaries are fairly firm but the lowest boundary with the P. asperopolus zone is chronologically very poorly constrained and not shown here.
oligotrophic environments but did not speciate during their austral excursions. Distinctive ecogroups have largely the same membership as biogeographical groups.

The larger foraminifers have expanded their latitudinal range at times of global warming and these forays have played the major role in reconstructing warming pulses in the Cenozoic (Hornibrook 1992; McGowran et al. 2000). The pattern theory of space–time distribution (Figure 8) is consistent with oxygen isotope and other palaeontological evidence for expansion of the warm-water zone.

**Macro-pattern of Cenozoic strata: the four-part subdivision**

The four-part subdivision of the Australian stratigraphic and biogeohistorical record is well-established. In Figure 9 it is shown as regional cycles I–IV, against which environmental and biotic events can be logged.

This is a second-order pattern and there are several ways of relating it to global schemes. The most obvious is the two-part Cenozoic expressed as the Palaeogene and Neogene Systems (even though the mid-Oligocene bottoming-out precedes the system boundary at the Oligocene–Miocene boundary). This pattern is seen more clearly as a climatic signal based in oceanic oxygen isotopes than in second-order sequences or regression–transgression cycles (Duval et al. 1998): indeed Haq et al. (1987) distinguished six of these and Hardenbol et al. (1998), eight. At this large-scale global pattern, the mid-Oligocene drop and mid-Miocene rebound are the main events on the 100 million years fall from the mid-Cretaceous to the present: the falling limb of the Mesozoic–Cenozoic cycle (Duval et al. 1998).

It is easy to see the chronological parallels with the four major (second-order) steps in the development of the late Neogene global environment (‘chills I–IV’). However, it is almost equally possible to make a case for tectonism as a significant control. Despite our increased knowledge of the Cenozoic Erathem, we do not yet understand the interplays between regional tectonism, tectono-eustasy, and glacio-eustasy. The curve of regional cycles in Figure 9 displays prominent lows at ca 10+, 30 and 44 Ma, and it is not difficult to find tectonic and climatic clues for these, as discussed below.

The best known stratigraphy and facies are in the southeastern corner of the continent, in Victoria. In two seminal papers, Bock and Glenie (1965) and Glenie (in Glenie et al. 1966) sketched the stratigraphic, transgressive–regressive cycles in the Otway Basin (Figure 10): the clearest antecedents at that time to the second-order cycles of today. Two major themes are the laterally extensive carbonate shelf facies (James 1997; Gallagher et al. 2003).
Sequence stratigraphy?

Becoming aware of the highly discontinuous record of the onshore neritic rocks, decreasingly enamoured of traditional biozonation in the southern mid-latitudes, and not

Figure 5 Cenozoic time-scales. Geochronology (Berggren et al. 1995) includes classical stages, magnetochronology (black, normal polarity; white, reversed polarity, chron C29–C1); radiometric calibration (Ma), and low- to mid-latitude planktonic foraminiferal zones (P, Palaeogene, M to PT, Neogene, as in Figures 3, 4). Third-order sequences and higher-order transgressive/regressive cycles (T/R), Hardenbol et al. (1998); Great Australian Bight (GAB) continental-slope hiatuses, Li et al. (2003b,c,d, 2004). Regional transgressions are good markers; regional stages (Chaproniere et al. 1996; McGowran et al. 1997a) are used sporadically in southern Australia, unlike the intensively employed New Zealand regional stages (Hornibrook et al. 1989).
wishing to abandon the regional chronostratigraphic stages, McGowran (1989a) perceived a natural division of the stratigraphic record based on rapid regional transgressions, too rapid for diachrony to be demonstrated biochronologically. This view of the record led to seeing a strong packaging of strata and their biofacies, and in turn

Figure 6 Southern Australia, with Cenozoic basins facing the Southern Ocean, in two climatic states based on the present (above) and the last ice age at 10^6 years scale (below). The Leeuwin Current hives off from the warm, low-salinity South Equatorial Current via the Indonesian throughflow, thus sampling the Pacific warm pool. During third-order and Milankovitch-order cool times the West Wind Drift and Subtropical Convergence shift north and the Leeuwin Current shuts down. The East Australian and Leeuwin Currents expanded the latitudinal ranges of essentially tropical larger benthic foraminifers, also warm and wet conditions thence vegetation and chemical weathering in southern Australia (McGowran & Li 1998; McGowran et al. 2000).
led to a regional version of the ubiquitous tension between lithostratigraphy and sequence stratigraphy. To illustrate this dialectical opposition, consider stratigraphic diagrams for the Port Campbell, Gambier and Murray Basins (Figure 12). Like numerous such reconstructions, it displays lithofacies and boundaries of formations (even members) persisting for millions and in one case tens of millions of years. Such diagrams are very different from allostratigraphic packages defined by unconformities and within which facies patterns are plausible. Contrast this prolonged diachrony with the coal–limestone package (cycle or synthem) in Figure 12: surely a more

![Diagram](image_url)

**Figure 7** (Above) Distribution and relative abundance of modern planktonic foraminiferal species (Bè & Tolderlund 1971; Bè 1977). (Below) Influence of the Leeuwin Current across the Great Australian Bight on foraminifers in neritic sediments. Benthos: species which disappear west-to-east include most notably four larger forms (*Amphistegina, Heterostegina, Planorbulinella, Amphisorus*). Plankton: the subtropical/temperate boundary meets southwest Australia at about the southwest corner (Cape Leeuwin) in large-scale maps but the Leeuwin Current actually takes the province onto the southern margin: the west-to-east changeover at ~125–127°E can be interpreted as a cooling which forces a weakening of the Leeuwin Current’s pelagic influence in the face of the Roaring Forties. From Li and McGowran (1998).
heuristic and challenging stratigraphic model? The coal-limestone package focuses attention on unconformities as powerful unifiers and packagers of marine-plus-non-marine strata.

**Sequence palaeobiology?**

Recent terms such as ‘sequence palaeobiology’ and ‘evolutionary palaeoecology’ refer to a reinvigorated and optimistic view of the fossil record as comprising something much richer than impoverished fossil versions of modern biotas. As well as the gradualistic and uniformitarian peering back into time, the fossil record can frame and test notions about communities and evolution that are simply inaccessible to students of the modern biota. It is evident that (i) the chronologically constrained environmental scenario of the Cenozoic is unique in the fossil record, and (ii) microfossils are the most informative fossils in the Cenozoic. Hence too, and more importantly, it is our plea that students of the rich records of several fossil groups in southern Australia take advantage of the chronologies and environmental scenarios now available to investigate community dynamics and evolution at long (supra-ecological) time-scales.

**CENOZOIC SEQUENCE I: PALEOCENE AND EARLY EOCENE**

We can identify three successional horizons or intervals in this part of the central-southern passive margin: accelerated divergence in the late Middle Eocene; Late Miocene basin inversion; and Late Neogene uplift.

The accumulation of sediments began as part of the onset of a major episode of accumulation along the entire continental margin. This event, of late Middle Eocene age (ca 42 Ma), was a response to: (i) plate rearrangement (Veevers et al. 1991) following India–Asia collision (Paleocene); (ii) accelerated Australia–Antarctica divergence; and (iii) marginal subsidence including either formation or rejuvenation of every sedimentary basin around and across Australia.

The southern and southeastern continental margins were shaped by strong structural trends in eastern Gondwanaland that go back to the Permian and early Mesozoic breakup, and subsequent dispersal of the continental fragments, including Australia. Oceanic crust was emplaced during the Cretaceous (beginning ca 99 Ma, Southern Ocean; ca 80 Ma, Tasman Sea) and ocean-floor spreading was very slow. Siliciclastic wedges were

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**Figure 8** Extratropical excursions by neritic, essentially tropical larger foraminifers. There were always larger photosymbiotic forms on the shelves, platforms and atolls of the tropical Indo-Pacific Province and dispersal into southern Australasia is well-corroborated. It is reasonable to question whether an episodic fossil record records episodic biogeographical events vs merely a sporadic fossil record. From McGowran et al. (2000) who inclined strongly towards the former view. Abbreviations at left: four warming events as in Figure 2.
accumulating in the Gippsland, Otway and Great Australian Bight Basins (Norvick & Smith 2001). Apatite fission track results indicate a major episode of uplift and denudation at ca 94 Ma, post-breakup (O'Sullivan et al. 2000), no doubt stimulating the delivery of sediments. Ocean-floor spreading ceased in the Late Paleocene in the Tasman and Coral Seas at the eastern continental margin, as components of a tectonic caesura including India–Asia collision, as discussed from an Australian perspective by McGowran (1989a, 1990) and Veevers (2000a), and from an ocean drilling strategy by Exon et al. (2002). McGowran (1989a, 1990, 1991) used a three-part tectonostratigraphic succession as the physical basis for a three-part environmental and biospheric succession at the global scale. The ‘old’ or ‘Laramide’ phase ended with the cluster of events around the Paleocene–Eocene boundary, especially India–Asia collision, which forced plate reorganisation. After an interregnum of some 8–10 million years the ‘new’ or ‘Himalayan’ phase began in the late Middle Eocene (ca 43 Ma) with the widespread manifestation of that reorganisation, including a change in Pacific motion and accelerated Australia–Antarctica separation (names of global tectonic regimes from Ziegler et al. 1985). Regionally, this latter phase is seen in continental deformation (arches, domes and monoclones, and faulting or rejuvenated faulting) and thermal subsidence causing the formation or rejuvenation of numerous sedimentary basins on the margins and in the continental interior. The major southeastern Australian Cenozoic episode of uplift and denudation was seen on fission track evidence to be at some time between ca 65 and ca 45 Ma (O’Sullivan et al. 2000), now more precisely linked with rapid spreading and thermal subsidence (Kohn et al. 2002). This will surely turn out to reflect the stratotectonic key number of ca 43 Ma.

Palynostratigraphy has established the stratigraphy of the Paleocene and Early Eocene in some detail. The sediments are siliciclastics, organics-rich but carbonate-poor. The facies are seen as comprising a spectrum from non-marine, to marginal marine with dinoflagellates accompanying the terrestrial palynofloras, to marginal marine with agglutinated foraminifers accompanying both, punctuated by relatively open-marine ingressions, named and ranging from the Ceduna (in the Maastrichtian: Figure 5) to the Burrungule (late Early Eocene: Figure 3). The marine ingressions are marked by calcareous benthic and sparse but valuable planktonic foraminifers, some molluscs, and rare shelly invertebrates.

There is some evidence that the marine ingressions have each its own distinctive foraminiferal fauna, ordinated consistently, and useful as chronostratigraphic markers interbasinally (McGowran 1991) (Taylor zones U–P in Figure 3). The faunas have not been investigated thoroughly enough, ceding to palynology in constraining the stratigraphy of these siliciclastic wedges. However, the ingressions fall in the time-slice of highest frequency of putatively global third-order sequences (Figure 3). Holdgate (1981) described a succession of packages in the Otway Basin siliciclastics on downhole physical criteria, interpreting them autocyclically as deltaic cycles. We expect that adequate sampling and biofacies study in due course will reveal each cycle to contain a marine ingress in the transgressive tract and that they will fit the sequences allocyclically. [Compare the number of regionally correlative shales, Pebble Point to Burrungule (Figure 13) with the number of ‘global’ third-order sequences in the same time-slice (Figure 3): the numbers are not dissimilar.]

**Lutetian gap and the Latrobe unconformity**

The loss of biostratigraphic (or any other chronological) control between ca 50 and ca 42 Ma is apparent in several

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**Figure 9** Environmental change and biotas in southern Australia. Left, two curves sketching the two-part Cenozoic approximating the Palaeogene and Neogene as two oceanic oxygen-isotopic cycles Pi and Ni (Abreu et al. 1998) as well as two putatively eustatic supercycle sets Tejas A and Tejas B (Haq et al. 1987, 1988). Amplitudes are probably exaggerated in the latter (Kominz et al. 1998). The four regional cycles are sketched to show: A, more global and regional environmental generalisations; and B, more regional biotic generalisations. Acronyms for four warming events as for Figure 2. From McGowran et al. (2000) and various sources.
figures (see especially Figure 3). In only one offshore section (the pelagic carbonates drilled by DSDP 364 on the Naturaliste Plateau) are the planktonic foraminiferal events between the Burrenugle and Wilson Bluff horizons seen in southern Australia (McGowran 1981). Thus the ages of the contacts between the Taylor zones N, O and P are poorly constrained; zone O is shown in Figures 3, 14 as higher (younger) than previously shown (see below). This uncertainty in marine micropaleontology exacerbates the central biostratigraphic problem in terrestrial palynology: the very poorly constrained age of the contact between the Proteacidites asperopolus zone and Lower Nothofagidites asperus zone. The contact or interface could fall anywhere between 49 and 44 Ma, and removing this uncertainty is the major requirement for progress in southern Australian Palaeogene studies.

This ‘Lutetian gap’ in the record is both a regional and a basinal problem. Regionally, we have known for some time that not only the circum-Australian but the circum-northeastern Indian Ocean stratigraphic records have this gap (McGowran 1977). Between the Ypresian and late Lutetian–Bartonian marine strata there is uncertainty: the sediments are missing or poorly dated. It is surely no coincidence that this gap is coeval with the tectono-stratigraphic interregnum between the Laramide and Alpine regimes in the general succession.

The regional manifestation in southeastern Australia of this gap pertains to the Latrobe unconformity, with the additional problem that all observations and interpretations, with one exception, are in the subsurface. In the Otway Basin a stratigraphic cross-section offers a relatively straightforward interpretation (Figure 13). The Dilwyn Formation, with its regionally persistent repetitive sand–shale pattern and Early Eocene age (see the youngest horizon, the Burrenugle Member, on Figures 3, 13) has been deformed and succeeded by the Nirranda Group (maximum age ~Wilson Bluff transgression in Figure 3) and Heytesbury Group (Late Oligocene – Miocene). In the more intensively explored and developed Gippsland Basin the seismic and stratigraphic patterns are more complex and controversial (Partridge 1976; Threlfall et al. 1976; Taylor 1983; Rahmanian et al. 1990; Bernecker & Partridge 2001; Johnstone et al. 2001; Holdgate et al. 2003a). The Latrobe unconformity is perceived variously as of second-order and third-order status in sequence-stratigraphic terms. At the broad end of the range, it refers to a

Figure 10 Stratigraphic diagrams for the Otway Basin across four decades. (a) Location of sections, wells and bores, and towns. (b) Panel diagram with three transgressive-regressive cycles identified by Glenie et al. (1966). Cycle I is Cretaceous; 2 is our Sequence I; Cycle 3 shows a strong split within the Oligocene falling at our II/III boundary; our late Neogene sequence IV is shown here by a Quaternary cycle. (c) Stratigraphic reconstruction based largely on palynological and planktonic foraminiferal determinations (modified after Gallagher & Holdgate 2000). Datum is at the II/III boundary of our Cenozoic sequences. Panels B–B and C–C are at about the locations of the north-south panels in (b). The I/II sequence boundary is between the Burrenugle and Sturgess Point Members; the III/IV boundary falls between the Miocene carbonates and Pliocene volcanics and sediments.
prolonged state of very low rates of sedimentary accumulation from the Late Maastrichtian to the Late Oligocene (~40 million years) due as much to non-deposition as to erosion (Bernecker & Partridge 2001). Also prolonged but more constrained was a hiatus seen as occupying most of the Eocene (Threlfall et al. 1976). Still more constrained would be the hiatus marking the Lutetian gap and tectono-stratigraphic interregnum. Most recently and most precisely, at the narrow end of the range in time and distinctly younger than the big gap, is an angular unconformity and hiatus restricted in concept and duration to ~2 million years, constrained by planktonic foraminiferal zonules O and N and itself constraining the ‘Latrobe uplift’ (Holdgate et al. 2003a). This concept of the Latrobe unconformity is shown in Figures 14–17. The problem of usage is partly a matter of definition but, more importantly, a matter of arguments about correlation. To take discussion of the Latrobe unconformity further necessitates discussion of the third-order hiatuses in the later Eocene (Li et al. 2003b), and they fall within the ambit of the second sequence to which we now turn.

CENOZOIC SEQUENCE II: LATER EOCENE TO EARLY OLIGOCENE

If there is a plausible natural beginning to geologically modern southern Australia, it is surely the Wilson Bluff transgression (~Lutetian–Bartonian boundary, ca 42 Ma). There are neritic carbonates (for the first time since the Palaeozoic) generated by well-skeletonised bryozoan-molluscan–echinoidal–foraminiferal biotas with, coevally, the sudden expansion of coal swamps along the southern margin extending from the far southwest to the southeast. Hence the stratigraphic notion of the ‘limestone–coal sandwich’ in the Late Palaeogene to Early Neogene in southern Australia (see below). These strata are found in sedimentary basins, all of which were generated or rejuvenated at that time, as the rather sudden acceleration of sea-floor spreading took place and Australia–Antarctica separation and birth of the modern Southern Ocean occurred. The Wilson Bluff is a regional component of the Indo-Pacific Khirthar transgression, recognised long ago as one of the most important stratigraphic events in that vast biogeographical region (Nagappa 1959), and considered to be some response to the ‘new’ or Himalayan regime, after India–Asia collision and after Australia–Pacific collision. However, cause and effect are unclear. For example, it is reasonable to ascribe the sudden onset of the regionally broad, neritic, carbonate-generating realm to regional causes such as ramping the margin just as Australia migrates into a more hospitable climatic belt. Sedimentologists, geomorphologists and terrestrial palaeobiologists have often argued that way. But the sheet limestones appeared (geologically) suddenly, not only on the southern continental margin in the...
Bartonian, but on the western and northern margins, with correspondingly warmer biotas, as well (McGowran et al. 1997a, b) (Figure 17). This correlation embracing the Wilson Bluff as a component of the Khirthar transgression ensures that biogeohistorical explanation will also encompass more than the Australian region. A better scenario than steady drift northwards into hospitable environments is to see those same environments moving further and faster southwards (Figure 1).

A sharp warming in the early Bartonian has long been indicated by these neritic and terrestrial data but is not apparent in oceanic oxygen isotopic series (McGowran et al. 1997a). However, Bohaty and Zachos (2003) recently demonstrated a sharp transient warming at ca 41.5 Ma over ~600 000 years, their ‘Middle Eocene climatic optimum (MECO’ (Figure 2).

Biostratigraphy

The planktonic foraminiferal succession summarised by Taylor (1986) has been important in the study of the Palaeogene succession in the Gippsland Basin. Taylor’s zonule P was based on the presence of Pseudohastigerina veloxensis and Planorotalites australiformis, suggesting earlier Middle Eocene (Figure 3), but the absence of Acarinina primitiva is plausible negative evidence for identifying it with the Burrungule planktonic assemblage (McGowran 1991). Zonule O is much younger; a change from its usual correlation (Holdgate et al. 2003a). Zonule O was characterised by the presence of Subbotina frontosa, ‘Globigerinatheka’ higginsi and Globorotalia centralis (assuredly Turborotalia pomeroli); all members of the Wilson Bluff assemblage (McGowran & Lindsay 1989 figure 2; McGowran 1989a figure 3). This assemblage is well constrained by the well-established first appearance of T.pomeroli (Figure 3), and we can take zonule O as approximating the Wilson Bluff assemblage. (The absence of Globigerinatheka index from the otherwise Wilson Bluff assemblage is unexplained.) Zonule N contains the last acarininids, thereby approximating the Tortachilla assemblage (McGowran 1989a figure 4), except for one species (Planorotalites renzi = pseudoscitula), an unexplained stray from below. Zonules K, L and M are not clearly characterised in terms of the succession of bioevents pieced together in Figure 3.

The terrestrial palynological succession is critical to arguments of Palaeogene succession and correlation. It is necessary to reaffirm the following correlations of the Nothofagidites asperus succession of three zones (Figure 3). Although it must have an interface with the Proteacidites pachypolus zone somewhere below, the Lower N. asperus zone is strongly associated with the strata and marine microfossils of the Wilson Bluff, with the Taylor zonule O in the Gippsland Basin, and with the Wilson Bluff foraminiferal assemblage found in the Otway Basin in the Sturgess Point Formation (Abele 1994). The Middle N. asperus zone includes pollen microfloras in strata at Browns Creek (A. D. Partridge pers. comm. 2003), strongly identified as including the sediments and microfaunas of the Tortachilla transgression and assemblage (McGowran 1989a). Thus the Middle/Lower N. asperus zonal boundary can be taken as being in the vicinity of the boundary.

**Figure 11** Sequence reconstruction in the Latrobe Valley and Gippsland Basin (Holdgate et al. 1996). The (Eocene) Traralgon T1 and T2 coal sequences are separated by the Latrobe unconformity (with low-angle deformation); the Mid-Oligocene Traralgon unconformity also shows deformation. Note too the dolomite marker at the top of both M2 Coal Sequence and lower Lakes Entrance Formation. The strong contrast of coal facies onshore with carbonate facies offshore developed in the Late Oligocene to Middle Miocene.
between the Wilson Bluff (and zonule O) and the Tortachilla (and zonule N). This has all been clear enough for some time (McGowran 1989a, 1991), but recently Holdgate and Clarke (2000), Hou et al. (2003a,b) and Clarke et al. (2003) have placed the Middle/Lower *N. asperus* zonal boundary too high: above the Tortachilla. The Upper/Middle *N. asperus* zonal boundary is at or just below the Aldinga transgression, a boundary once placed in the latest Eocene but actually just above the base of the Oligocene (McGowran et al. 1992), although the obsolete age determination lingers in Holdgate and Clarke (2000) and Clarke et al. (2003).

Bartonian–Priabonian marine transgressions, hiatuses and allostratigraphic packages

Patiently assembling the microfossil record slowly revealed that it was packaged as successional assemblages, each characterising a regional transgression (McGowran 1989a). These transgressions were rapid with no evidence of diachrony at the available resolutions, and named after prominent rock units (Figures 3, 17). The five named here (Wilson Bluff, Tortachilla, Tuketja, Tuit and Aldinga) are found in more than one basin and bear evidence of unconformity and hiatus. They have been found useful in

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**Figure 12** Two contrasting worldviews of stratigraphy. One is the perception of diachrony to stratigraphic reconstruction, with the zigzagging ‘Shazam!’ symbol (Brett 2000), traversing millions or even tens of millions of years as seen in many stratigraphic diagrams (including the authors’); in the other, meaningful diachrony is restricted to within-sequence facies boundaries (Holland 1999). (Above left) Murray Group carbonates and Renmark Group siliciclastics (with lignites) as visualised by Brown and Stephenson (1991). Although the vertical scale is time, intragroup unconformities and the ‘mid-Tertiary hiatus’ display no hiatuses (a measure of the very broad time constraints in those facies and the gradualist assumptions of the compilers). Likewise, the Renmark Group includes the siliciclastics below the ‘mid-Cenozoic hiatus. Paleocene–Eocene siliciclastics omitted. (Above right) Gambier and Port Campbell Basin lithostratigraphy from various sources (Li et al. 2000) with strongly diachronous boundaries throughout. Gambier and Port Campbell are located in Figure 10. (Below left) The Gambier Limestone (in the Niranda Group) in sequence terms. Stratal units 1–7 are sediment packages consistent with the third-order sequences of Hardenbol et al. (1998). Limestones and dolomites are the major lithologies in this borehole (RR65); third-order sequences and stratigraphic units were based on erosional surfaces, abundance changes in planktonic and deep-water benthic species, and clustered assemblages (Li et al. 2000). (Below right) Third-order coal-limestone cycle or synthem (Holdgate & Gallagher 1997). TST and HST, transgressive system tract and highstand system tract. The bar/hinge (here the Balook Sands) is prominent in all major accumulations. That the model is useful for the brown coal deposits in southern Australia is corroborated by correlations and comparisons between coals and limestones and between both and a curve of global ice volume (Figures 11, 26); and coals vis-à-vis large benthic foraminifers in identified sequences (Figure 23)
correlation and age determination notwithstanding the obvious risk of circular reasoning. By correlation, the unconformity-bounded stratal packages can be compared with putatively global third-order sequences (Hardenbol et al. 1998). We lack the means to regionally test the latter, but we can seek consistency and the match is good: five rapid transgressions fit five of the six available transgressive sequence tracts. [The temporally brief Tuit is both clear and consistent with the Hardenbol et al. (1998) succession, and marks an unnamed glacioeustatic horizon (Miller et al. 1998). However, it lacks its own planktonic assemblage; it lies immediately beneath the erosional downcut associated with glaciation O1 and its elusiveness accordingly is preservational. The Tuit is firmly established in the St Vincent and Otway Basins.] The only apparent inconsistency is at the Middle–Late Eocene boundary, where sequence boundaries Pr1 and Pr2 are one too many. However, the Late Bartonian Tortachilla Limestone has multiple hardgrounds (James & Bone 2000), one of which plausibly indicates SB Pr1 and a starred sequence of more eutrophic biofacies than the main Tortachilla.

Li et al. (2003a, c) confirmed the notion of foraminiferal biostratigraphic packages, one each for the Wilson Bluff, Tortachilla, Tuketja–Tuit and Aldinga, and found that four of the hiatuses could be traced into the Great Australian Bight, where previous inferences of rapid marginal subsidence associated with the new tectonic regime at ca 43 Ma (McGowran 1991) were substantiated. Hence, stratal packages extended from onshore to offshore (the below-Tuit hiatus is not seen, but it was vulnerable to sampling failure and Early Oligocene downcutting). This succession of packages spans the onset of ‘partial or ephemeral ice-sheets’ (Zachos et al. 2001b), an inferred event correlating with hiatus C, SB Pr1, and the Bartonian–Priabonian boundary at ca 37 Ma. Skeptical that glacioeustasy could control unconformities across the neritic-bathyal range of water depths via, say, slumping, Li et al. (2003a) developed a scenario of pulsed or stepwise increases in ocean-floor spreading in the early stages of the new regime. Thus, the onset of multiple transgressions clearly was a tectono-eustatic event at the second order; the multiples themselves in this transition to icehouse were also tectono-eustatic at the third order.

This work in the central-western part of southern Australia is consistent with the above-mentioned constraining of the Latrobe unconformity in southeastern Australia. Holdgate et al. (2002) identified a ‘mid Eocene tectonic event’ in the coal-bearing sections of the Gippsland, Port Phillip and Torquay Basins at the Middle–Lower Nothofagidites asperus zone boundary (Figure 3) and developed their arguments further (Holdgate et al. 2003a) for the Latrobe unconformity sensu stricto, tightly constrained as shown in Figure 16. Offshore in the Gippsland Basin, the Latrobe unconformity is directly below the

Figure 13 Stratigraphic section in the Otway Basin, located to the west of section C–C’ in Figure 10a, emphasising the thick siliciclastics in the Early Palaeogene not emphasised in the latter. The Pebble Point and Burrungule foraminiferal faunas bracket this Thanetian–Ypresian wedge (see marine ingressions correlated in Figure 3). Eleven regionally correlative shales are shown, inviting comparison with the ~12 third-order sequences for the same time-slice in Figure 3 (where the five named marine ingressions are only the relatively well-understood events). Unconformity ‘2, Middle Eocene’ combines the deformation and truncation associated with the Latrobe unconformity and Marlin unconformity in Gippsland (Figure 14), the latter also in the Torquay Basin (unnamed at A/B in Figure 15). Deformations were continental margin responses to the ca 42 Ma tectonic event.
Gurnard Formation (zonule N) and cuts across channel fills with zonule O assemblages. However, there is a case for perceiving multiple deformations chronologically separate from the Latrobe unconformity. Below the latter is the gap that surely is associated with the acceleration in ocean-floor spreading and the 43 Ma event. Above it is evidence of volcanism, such as the K–Ar date of 36.5 Ma in the Otway Basin (Figure 10e) at hiatus C (Figure 16). In the onshore Gippsland Basin in the Latrobe Valley coalfields (Figure 11), the Traralgon sequence including the Latrobe unconformity itself is deformed in the Early Oligocene. Thus the unconformity is one of several responses to the onset of the new tectonic regime: uplift, denudation and multiple transgressions.

Chemofacies and the Auversian facies shift

One of the most pronounced shifts in environments was at the Eocene–Oligocene boundary. Before (Late Eocene), the oxygen isotopic record from the deep ocean suggests that Antarctic icecaps were smaller and transient (‘the ice-house cometh’: Browning et al. 1996). After (Early Oligocene), icecaps were larger and relatively permanent (Zachos et al. 2001a with references). This shift was inferred to mark the origin of the psychrosphere (the cold mass of the ocean below the thermocline), and to that extent signals the beginning of the modern ocean and global environment. A major drive for this change was the opening of the Southern Ocean with acceleration of ocean-floor spreading (Middle Eocene, ca 43 Ma), and the first oceanic throughway on ocean crust south of Tasmania (Early Oligocene, ca 34 Ma). Environments and biotas responded to this global Eocene–Oligocene boundary shift in the pelagic, neritic and terrestrial realms in what is known as the Terminal Eocene Event (Prothero & Berggren 1992; Prothero 1994a, b). The main change was, variously, prior to and at the Middle–Late Eocene boundary and at the Eocene–Oligocene boundary. There is a strong sense of punctuated biotic change rather than of gradual change (and perhaps sharpened taphonomically: i.e. an artefact of the fossil record).

The later Eocene (Bartonian–Priabonian) world was more warm and humid, oceanic environments were poorly partitioned in slower circulation, chemical sediments were diverse and mixed, and poor ventilation was more common. By Rupelian time oceanic circulation was invigorated, sedimentary environments were more cleanly partitioned, and the world was ice-cap-prone. Berger and Wefer (1996) outlined these generalisations in naming the Auversian facies shift and identifying a fundamental shift in global oceanic environments at ca 40 Ma (McGowran & Li 2000). Until that time, pelagic carbonates were restricted to shallower oceanic depths and commonly mixed with cherts; subsequently, they were purer as well as substantially deeper (and therefore more widespread on ridges and plateaus), after the carbonate (calcite) compensation depth plunged by a kilometre and more. With the development of the psychrosphere came an increased efficiency in oceanic ventilation and a more effective partitioning of chemical and especially biogenic sediments. The transition was from a warm-ocean mode of sedimentation to a cold-ocean mode, and it was forced by oceanic closures at low latitudes and the opening of gateways at high latitudes: ‘the thermohaline system moved slowly but inexorably from a Warm-Ring-dominated world toward the present configuration of a central [circum-Antarctic] Cold Ring’.

Figure 14 Stratigraphic diagram for the Palaeogene, Gippsland Basin, modified after Bernecker and Partridge (2001) and Holdgate et al. (2003a). Three problems are shown: (i) biostratigraphic resolution and constraint are poor in the early Middle Eocene (the ‘Lutetian Gap of Figure 17) for both palynology and planktonic foraminifers (Taylor zones); (ii) this biostratigraphic shortcoming causes uncertainty in position of the Marlin unconformity; the younger of the two levels shown is from Bernecker and Partridge (2001); and (iii) a pronounced temporal range in notions of the Latrobe unconformity is shown by two end-members; long duration, Bernecker and Partridge (2001); short duration, Holdgate et al. (2003a) and this paper. Ch, channel; U/C, unconformity; Fmn, Formation.
The Auversian facies shift is a bold generalisation about a pervasive change in state in the global ocean closely following the tectonic changes at ca 43 Ma (and also responding to the same: see McGowran & Li 2000).

Facing this emerging cold ring, the southern Australian continental margin recorded in its neritic environment the various stages of the Auversian facies shift: incursions of benthic warm-water organisms during warmings and high sea-levels, and their absence during coolings and regressions (the plankton parallel the benthos with a clearer cool/fertile signal). The oceanic generalisations captured by the Auversian facies shift are paralleled remarkably in the neritic realm. The onset of the new tectonic regime stimulated a transgression and neritic carbonates, some cherty (Wilson Bluff) and some oligotrophic (Tortachilla) and some basal glauconites. The Late Eocene, by contrast, records rather diverse facies with a signature of rather poor ventilation: greensands, opaline silicas, dark-grey marls, muds and sands with biofacies characterised by high productivity and strongly infuonal microbiotas and macrobiotas. The silicas have been studied as to their opal and clay mineralogy (Jones & Fitzgerald 1986), biofacies and correlation (McGowran & Beecroft 1986) and sedimentology (James & Bone 2000) at Maslin and Aldinga Bays in the St Vincent Basin, the best exposed and best known of the opaline silicas of west-southern Australia (see also Gammon et al. 2000, 2003a, b). A recent generalisation about Late Eocene neritic facies has it that there is a north–south lateral gradation along 2000 km of the continental margin from inshore (and palaeovalley) opaline silicas to offshore (still neritic), bryozoan carbonates (Gammon et al. 2000, 2003b). However, the generalisation is unsupported: the only stratigraphically constrained neritic carbonates of Priabonian age are in the Otway Basin (Figure 16). The pattern is not lateral and coeval but temporal and successional: earlier Bartonian chert–carbonates (Wilson Bluff) and coals; then later Bartonian oligotrophic carbonates (Tortachilla); then Priabonian coals and diverse chemical facies (Tuketja–Tuit). Most of the Late Eocene diversity is swept away by the changes ushered in on glaciation Oi1; the Rupelian (Early Oligocene) neritic in southern Australia is typified by carbonates and sands on a well-ventilated shelf. The spectacular contrast shown in the quantified biofacies of benthic foraminifers in one section (Figure 18) faithfully records in microcosm the contrast across the Eocene–Oligocene boundary (see below).

The Late Eocene and Oligocene strata are clearly allostratigraphic packages that are interbasinal and traced across the shelf/slope break in the Great Australian Bight.

Later Palaeogene coals

Some of the most extensive coals in the global Cenozoic record began accumulating in the Bartonian (Holdgate & Clarke 2000; Holdgate et al. 2000b). Right along the southern Australian margin there is recurring evidence that the coal swamps occurred in the classical fashion, close to the shoreline but in grabens, palaeochannels or back barriers. They occurred in three pulses, comprising part of the allostratigraphic packages associated respectively with the transgressions named Wilson Bluff (Lower N. asperus zone, Bartonian), Tuketja–Tuit (Middle N. asperus zone, Priabonian), and Aldinga (Upper N. asperus zone, Rupelian), and confined to the same third-order sequence in each case. Interestingly, the most widespread carbonate-generating transgression, the Tortachilla, seems to be unaccompanied by coals (possibly excepting Gippsland), an interesting anomaly in the configurations in the ‘coal–limestone sandwich’. Using Neogene examples, we have shown that coals occupy the same place in third-order sequences onshore that limestones occupy offshore, that is, concentrated in the highstand (Holdgate & Gallagher 1997) (Figure 12). This pattern holds in the Palaeogene coals of the Traralgon field and others in southeastern Australia. Holdgate and Clarke
(2000) suggested that the Late Palaeogene coals in the western parts of the margin depart from the model by occurring in the transgression. However, this is probably not so in the St Vincent Basin, and the correlations in the Bremer Basin are not refined or robust enough to sustain the exception to the Neogene sequence-pattern.

Episodic deep weathering?

The deep weathering characterising so much of the Australian landscape is of Paleocene–Early Eocene age, the warmest and most humid times of the Cenozoic when southern Australia was in a rainbelt, an argument developed by McGowran (1989b) to explain the exhalation of silica in oceanic facies, manifested in the cherts producing the interoceanic seismic Horizon-A. There are two perceptions of deep weathering on Australia (and east Gondwanaland): gradualist and episodic (‘neocatastrophic’?). For Lyellian gradualists, weathering goes on all the time and the spatiotemporal variations in intensity are neither clearly identifiable nor particularly interesting. In the most recent study in this mode, Taylor and Shirliff (2003) assembled a wide range of regolith age determinations into histograms by coarse time-slices, Permian–Neogene, and failed to disprove a null hypothesis that weathering intensity shows no marked fluctuation through time. In a discussion in the alternative, episodic paradigm, McGowran and Li (1998) attempted to show that all other components of the global exogenic system have stop-and-start patterns and thresholds (e.g. ocean-floor spreading systems and orogenic pulses, palaeoclimatic–palaeoceanographic shifts, and punctuated organic evolution), and it would be strange indeed if the regolith were exempted from this pattern. They posited that the four second-order truncated warmings of the Cenozoic Era (Figure 2) were also truncated episodes of deep regolith weathering and ferruginisation (McGowran et al. 1997a). Unfortunately, neither argument had sufficiently refined chronological data to decisively advance the debate. We lack records of ordination and succession in the regolith comparable to the step-by-step mineralogical and pedological changes through thick sequences in North America recording the Eocene–Oligocene environmental shifts (Bestland et al. 1997; Bestland 2000), where bundles of palaeosols bounded by erosional surfaces in fluvial sediments have been regarded as the terrestrial equivalents of marine allog stratigraphic packages.

However, Pillans (2002; also McQueen et al. 2002) has tested the two models of weathering, continuous and episodic, with palaeomagnetic dating of iron oxides. There is a marked tendency for dates to cluster, with one ‘particularly remarkable’ cluster at 60 ± 10 Ma and a second at 10 ± 5 Ma (Figure 2), offering strong evidence for an episodic pattern. Further, in oceanic data on clay mineral distribution on the Maud Rise and Kerguelen Plateau (Ehrmann & Mackensen 1992; Salamy & Zachos 1999; Zachos et al. 1999), only a few degrees palaeolatitudinally south of southern Australia, four oceanic sections show a dramatic switch from dominantly chemical weathering (on the evidence of smectite and kaolinite) to dominantly physical weathering (illite and chlorite), the transition occurring in the Late Eocene. This chronological pattern is powerful support for one of our postulated truncations of deep regolith weathering, that at the end of the Eocene. Wopfner and Walther (1999) emphasised that the really extensive silcretes comprising a large part of the carapace of continental Australia accumulated on the Cordillo Surface and were of later Eocene age, their accumulation stimulated in the first place by the change in continental environment in the later Middle Eocene. Why, then, did Paleocene–Eocene deep weathering result in deep-ocean silica (McGowran 1989b), whereas later Eocene deep weathering and ferruginisation (McGowran et al. 1997a).
weathering resulted in extensive silcretes (quite possibly but not demonstrably coeval with the neritic opaline silicas)? The answer would seem to be in the formation of endorheic drainage by continental warping as still another outcome of the ca 43 Ma caesura in Australian geodynamics.

The regolith mantle was cut by Early Eocene channels in several basins, but much more extensive cutting originated at, or was rejuvenated by the tectonic event at ca 43 Ma. The palaeochannels are best known in the Gippsland offshore oil and gasfields, the Gippsland coalfields, and the Victorian deep leads. In the offshore Gippsland Basin the Latrobe unconformity overlies infilled channels, probably marine canyons, in the top of the Tuketja Group with zonule O foraminifers, prior to the flooding in zonule N. In the Bremer and Eucla Basins there are non-marine and marginal-marine sediment-filling channels (Hou et al. 2003a,b). The channels were flooded by the sea twice: once at the Tortachilla transgression and once at the Tuketja transgression. In the St Vincent Basin the channels became estuarine, flooded brackishly on the Wilson Bluff transgression. The channel fills in their turn were weathered and lateritised in the Eocene. The youngest set of channels under reasonable chronological constraint in southern Australia are at the Eocene–Oligocene boundary, backfilled before the Aldinga transgression. This is not to deny the existence of Oligocene downcutting, but this is not cogently dated.

Foraminiferal biofacies across the Eocene–Oligocene boundary, St Vincent Basin, southern Australia

The pivotal boundary horizon is a downcut (Chinaman Gully downcut) of up to 50 m and a siliciclastic backfill followed by a marine transgression found on both sides of the St Vincent Basin (McGowran & Li 1997b). Microfossil correlations are consistent with this event being the local manifestation of glaciation Oi1 (Figure 18). There is strong correlation of this section with the third-order sequences of the Bartonian, Priabonian and Rupelian Stages. The major biofacies contrasts across the Eocene–Oligocene boundary are seen in the dominant families of benthic foraminifers, the epifaunal Cibicididae and infrafaunal Uvigerinidae. There are three outstanding features: (i) the decrease in infauna; (ii) the dampened amplitude of swings; and (iii) the most pronounced taxic overturn through the Priabonian and Rupelian (not shown in Figure 18). Thus there is a very strong local neritic signal to one of the more significant global transformations during the Cenozoic Era: the onset of well-established ice sheets and the development of the psychrosphere. The changes at the Eocene–Oligocene boundary are wide-ranging: opaline to quartzose; infrafauna-dominated to epifauna-dominated microfaunas; grey-green to yellow-brown colours; sponge-rich to bryozoan-rich macrofaunas; broadly, from somewhat poorly aerated...
There were several well-marked mineralogical changes within the Late Eocene opal-rich section. The abrupt change from opal-CT to opal-A is at the top of the darkest sediments richest in infaunal gastropods (Spirocolpus) and trace fossils (Thalassinoides), exactly at a diagenetic change from hard–soft couplets to soft spicular marls and securely identified as the maximum flooding surface (Gull Rock/Perkana boundary). This is the strongest change in biofacies, especially in the relay from Uvigerinidae to Bolivinidae in the dominant micro-infauna, which is reversed at the abrupt reversal from opal-A to opal-CT and reversion to hard–soft couplets, identified as the next sequence boundary and the Tuit transgression. The environment was restricted in circulation and planktonic numbers were very low. The sequence boundaries are well-marked but hiatuses are brief and constrained. It seems likely that the patterns shown are not greatly affected by the biases on the fossil record imposed by sequence-stratigraphic architecture. The sequence architecture clearly controls the alternation of opal-A and opal-CT in parallel with infaunal dominances (especially Uvigerinidae–Bolivinidae), suggesting that the controls are in nutrient variations.

The contrast between the mixed facies of the Late Eocene and the carbonates of the Early Oligocene holds in the St Vincent and Otway Basins, especially the Gambier Limestone in the west, and also in one section drilled west of Tasmania on ODP 189. However, the glaciation Oi1 did not terminate some prominent Late Eocene characteristics. There was a warm-water rebound seen immediately in southern Australia and New Zealand (see the Halkyardia rebound in Figure 8), which is consistent with an oxygen-isotopic rebound after Oi1 (Zachos et al. 1996). Also, the environmental conditions conducive to massive coal accumulation at the ‘apogee of a greenhouse world’, warm-temperate to subtropical climate and highstands (Holdgate & Clarke 2000; Holdgate et al. 2000b), were disrupted only briefly by Oi1.

**CENOZOIC SEQUENCE III: LATE OLIGOCENE TO MIDDLE MIocene**

The true bottoming-out of the Eocene second-order cycle is not the first major glaciation (i.e. Oi1) but the Middle Oligocene (Figure 9). This is seen in the sequence-stratigraphic, long-term, Ypresian–Rupelian regression (Hardenbol et al. 1998), as well as in the oxygen-isotopic indications that Antarctic ice-sheets persisted through most of the Oligocene (Zachos et al. 2001b). It is also seen in southern Australia: the next second-order cycle begins with the pronounced Chattian warming, seen more clearly in recent studies (Zachos et al. 2001b), and seen regionally in the molluscan faunas (Darragh 1985; McGowran et al. 2000) and the first large-foraminifer ingression of the new order (Figure 8). Although the Neogene begins

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**Figure 18** Foraminiferal biofacies across the Eocene–Oligocene boundary, St Vincent Basin, South Australia (McGowran et al. 1992) with unambiguous identification of the Hardenbol et al. (1998) sequence nomenclature. The changeover from the essentially infaunal Uvigerinidae to epifaunal Cibicididae across the Eocene–Oligocene boundary (reinforced by the infaunal index) parallels a change from grey-green opaline chemofacies to typically inner neritic bryozoan–quartz facies. The Chinaman Gully downcut attains ~50 m in the St Vincent Basin backfilled by marginal marine siliciclastics and the upward-deepening Aldinga Member. The opaline chemofacies are distributed not merely diagenetically (James & Bone 2000) but stratigraphically: opal-CT in the transgressive systems tract and opal-A in the highstand systems tract (HST), also paralleled by pronounced biofacies changes (Uvigerinidae, Bolivinidae, Miliolidae and Cassidulinida). These clades track environmental perturbations in this nutrient-rich environment more succinctly than do the clades in the Early Oligocene. SB, sequence boundary; MFS, maximum flooding surface.
chronostratigraphically at the deep but transient glaciation M1 (Steininger et al. 1994; Shackleton et al. 2000; Zachos et al. 2001a), biogeohistorically it begins within the Chattian.

This time-slice, the upper half of the ‘limestone–coal sandwich’, is marked by great coal swamps in Gippsland together with the most extensive distribution of neritic carbonates in southern Australia (Figure 19), in what has been called a cool-water carbonate depositional realm (James 1997). The neritic carbonate seas extended their reach during the Oligo-Miocene to a maximum in the early Middle Miocene, most strikingly in the Eucla and Murray Basins. The largest actual mass of carbonate in the region is the Seaspray Group including the Gippsland Limestone, recently studied seismic-stratigraphically (Feary & Loutit 1998; Holdgate et al. 2000a) and in terms of biofacies and environments (Gallagher et al. 2001). In contrast to this massive limestone body, under strong oceanic influence including upwelling, are the carbonates of the Murray Basin, much thinner but laterally extensive and accumulated in an inland sea largely shielded from oceanic influence by the Padthaway Archipelago, islands of Delamerian (Early Palaeozoic) granites (Li & McGowran 1999; Lukasik et al. 2000). For this situation, Lukasik et al. (2000) developed the notion of the ‘epeiric ramp’, sharing some characteristics with the carbonate ramp and some with the epeiric seas of the great Palaeozoic and Mesozoic transgressions. The horizontal scale is hundreds of kilometres, water depth was a few tens of metres.

The Oligocene and Miocene carbonates in the Otway Basin are sheets, some tending to coarsen westwards from marls to grainy limestones (Figure 10). Sequence packing is illustrated for the Gambier Basin (Figure 12). Dolomitisation is locally strong, especially in the Oligocene: a mid-Oligocene horizon is found in several basins as well as in the open ocean (McGowran et al. 1997a) and a dolomite marker is a strong stratigraphic control in the Gippsland Basin at this same level (Figure 11). Kyser et al.

**Figure 19** (Lower) Palaeogeographical sketch of the southeastern part of the southern Australian margin in the early Middle Miocene at the apogee of marine transgression and carbonate facies (MICO in Figures 2, 8, 9) from Brown and Stephenson (1991), Alley and Lindsay (1995) and others. There is consensus that these shores were rimmed by rainforests, continuously or as a large proportion of the vegetation mosaic. Mt Lofty–Flinders and Olary Arcs were probably low hills. (Upper) Contrasting cross-sections across the present Lacepede Shelf, at present-day sea-level with a curve for 0–100 000 years (Li et al. 1996), and the western Murray Basin with a curve for the Miocene, 20–10 Ma interval using the Haq et al. (1988) curve (height is exaggerated: Kominz et al. 1998). Formations reflecting lithofacies belts: Brown and Stephenson (1991).
(2002), building on James et al. (1993), reconstructed a geochemical sequence of dolomitisation and dedolomitisation through the Cenozoic carbonates of the Gambier Platform (or Embayment), diagenetic episodes having some possible matches with transgressions, but found no strontium ages surviving from this Oligocene initiation.

**Jan Juc and Neogene packaging**

The Late Oligocene succession in the Torquay Basin invites a regional test of consistency with the putatively global third-order cycles and this consistency is sustained into the Miocene (Figure 20). The proxy for eustasy is the inner-to-outer benthic foraminiferal ratio, with shallowing predicting a sequence boundary and a third-order glacial (at least in ice-sheet-prone times).

Planktonic foraminiferal biostratigraphy on ODP leg 182 drilling in the Great Australian Bight (Li et al. 2004) revealed 15 hiatuses, each characterised by the absence of at least part of a biozone or subzone, and/or an unconformity, hardground or slump, which coincide with third-order sequence boundaries. In the Early Miocene, five hiatuses are correlated with five of the seven sequence boundaries. In the Middle to Late Miocene, six of the seven sequence boundaries are represented by hiatuses under the Great Australian Bight. Three of the four Plio-Pleistocene hiatuses identified seem to conflate more than one of the sequence boundaries, which are more frequent here than earlier in the Neogene. Given the less-than-perfect chance of complete records and completely successful sampling by drilling, these results are good evidence of regional concurrence with putatively global third-order sequence boundaries. Quite independent of this work was a biofacies curve based on the inner/outer neritic ratio of benthic foraminifers in a section in the Gippsland Basin, this ratio giving a proxy for a local sea-level curve (Li & McGowran 1997, 2000). Shallowings signalled sequence boundaries in most cases. Thus we have different lines of evidence in the Neogene, plus the later Palaeogene evidence already discussed above, that the strong pattern of regional stratal packaging reflects the global order.

**Coal–limestone couplets at the third order**

The coals and coal–limestone stratigraphic relationships were reviewed by Holdgate et al. (1995), Holdgate and Gallagher (1997) and Holdgate and Clarke (2000); the coal-carbonate configurations in the Gippsland Basin are shown in Figures 21–23. The emphasis put on allostratigraphic packaging is obvious, as is the remarkably stable position of the repeated sand bodies comprising the Balook Formation at the coal–limestone fulcrum in the Gippsland Basin. The third-order patterns are twofold on either side of the hinge sands: more marly carbonates coeval with brackish-marine clays tend to be in the transgressive tract; limestones matching coals tend to be in the highstands (as shown schematically in Figure 12, lower right). We emphasise that allostratigraphic packages pervade this stratigraphic record and infer that their absence is a function of insufficient data. We show one example of a downcut and sequence boundary in the Yallourn Coals (Figure 24) that might correlate as shown in Figures 23, 25.

The coal–limestone sequence-couplet implies a wide range of trophic resources. Increased warmth implies increased precipitation and increased production and accumulation of plant material. In contrast, the neritic carbonate environments host the large, symbiont-bearing, benthic foraminifers: excellent indicators not only of pronounced warming but also of oligotrophy. Hence the expansion of the trophic resource continuum at both ends: eutrophy and oligotrophy. This can be tested by plotting the stratigraphic distributions of neritic warming indicators.
Figure 21 Coal-carbonate sections, terrestrial–marine transition, Gippsland Basin (modified after Holdgate & Gallagher 1997 Figure 3). The schematic, third-order, coal-limestone sequence in Figure 12 was based on this reconstruction, which shows much of the detail omitted from the onshore–offshore facies patterns in Figure 11.
and coals against Cenozoic time and a putative curve of global ice volume. At the long time-scale (Figure 26) there is a strong correlation of warming, immigration by oligotrophic foraminifers and the accumulation of coals: the mid-Cenozoic, second-order ‘limestone–coal sandwich’ is clear. At the third-order time-scale (Figure 23), the couplet seems to hold for some sequences and there is some indication that the third-order couplets are responses to ice-sheet decay.

**Miocene warming to Middle Miocene optimum and its termination**

We have known since the 1950s (Durham 1950; Dorf 1955) that the Middle Miocene marks a warming to accompany maxima in: (i) the extent of transgressions across the continental margins leaving their mark in limestones and warm-water fossils; (ii) the latitudinal expansion of neritic and planktonic biotas; (iii) the extent and volumes of coals; and (iv) the extent of preserved strata of terrestrial facies (Hornibrook 1992; McGowran & Li 1994; Flower & Kennett 1995; McGowran et al. 2000; Gallagher et al. 2001).

The brown coals of the Latrobe Valley, up to 200 m thick, host microfloras and macrofloras rich in signals of Miocene palaeoclimates (Blackburn & Sluiter 1994; Kershaw et al. 1994; Holdgate et al. 1995; Sluiter et al. 1995; Gallagher et al. 2001). Sluiter et al. (1995) concluded that mean annual temperatures of 19°C are indicated for most of the coals, 4–5°C more than today; and mean annual rainfall probably of 2000–2200 mm, almost twice today’s; and the vegetation type was classed as subtropical rainforest at this palaeolatitude of ~55°S. The evidence for wet-subtropical conditions was strongest in the Morwell 1A and Yallourn coals, matching nicely the evidence of the neritic larger foraminifers (Figures 23, 26). Gallagher et al. (2001) distinguished a late Early Miocene warm
phase and Middle Miocene climatic optimum, respectively
Longfordian/Burdigalian (and Morwell 1A coals) and
Balcombian/Langhian (and Yallourn coals), and separated
by a sharp stratigraphic discontinuity marking glaciation Mi2. These are the last two third-order warmings of
several, beginning with the pronounced warming in the
Late Oligocene (Zachos et al. 2001b) and separated by third-order glacials.

All of the above indicators point to a pronounced environmental shift from warm-wet transgression to cool-dry regression, proceeding from the Middle Miocene into the Late Miocene across what is labelled as ‘chill III’ in
Figure 2. Oceanic oxygen isotopes support this generalisation (Shackleton & Kennett 1975; Miller et al. 1987, 1998; Zachos et al. 2001b) and it extends with environmental impact on the biotas in the pelagic, neritic and terrestrial environmental realms (McGowran & Li 2002). The critical time of this shift is Langhian–early Serravallian (Figure 25). The correlations in this time-slice must be distinguished from the numerical ages; the latter may well be revised cyclostratigraphically and the former surely can be improved. There are a variety of correlations here: standard and regional biostratigraphy, sequence stratigraphy, and chemostratigraphy. Note the relatively high frequency of the chemostratigraphic peaks: the cluster of oceanic carbon positives (CM1-6, later, CM7) and oxygen negatives (A-F) punctuated by the oxygen positives (third-order glacials Mi1b-Mi5).

We have assembled an array of foraminifer assem-
blages and stratigraphically-packaged events in southern Australia (Figure 25) to demonstrate that: (i) there are three bounding chronosurfaces that hold true through geography, environmental realms and facies; and (ii) this pattern is consistent with the putatively global succession. The biogeographical and nutritional pattern is twofold in the two fossil groups that are relatively well-constrained stratigraphically. One group is the large benthic foramin-
ifers, based on the stratigraphically constrained taxonomy and distributions by Chaproniere (1975) and discussed stratigraphically by McGowran (1979). These taxa spread around southern Australia and New Zealand in two waves in the Batesfordian–Balcombian (=early Langhian–
early Serravallian; Clifdenian–early Lillburnian in New Zealand) signalling two warm-oligotrophic peaks. The other group is the planktonic foraminifers displaying a cancellate/spinose ratio with two strong peaks also signalling two warm-oligotrophic peaks. Thus, we see a between-realm, facies–bioeogoeographical pattern that is particularly volatile or metastable at that time (McGowran & Li 1996).

The Middle Miocene (Serravallian) event marked globally by sequence boundary Ser2 and glaciation Mi3 (= chill III) is marked regionally by four sets of observations and correlations: (i) there is a surface on the carbonates in the Murray, St Vincent and Eucla Basins; the Nullarbor Plain and the surface in the Murray Basin are

Figure 23 Two benthic curves from the Lakes Entrance Oil Shaft (Li & McGowran 2000), one (left) an epifaunal/infaunal ratio (% infauna), the other (right) an inner/outer neritic ratio (palaeodepth estimated, a proxy for sea-level) Both are consistent with: (i) the ages of third-order sequence boundaries (Hardenbol et al. 1998); and (ii) the third-order glacials identified on oceanic δ18O signals (Mill-
er et al. 1998). Within this framework coals and large benthic foraminifers go together, as do coals from the two hemispheres (Lower Lusatia, Germany: Diessel 1998). Chronological matching of coal deposits with the two major, positive carbon-isotope excursions of the Miocene implies cause and effect: Australian coals clearly contributed very significantly to the Monterey carbon excursion, named after Californian and other northeast Pacific oil sources.
that surface, suggested long ago (Glenie et al. 1966) to be a glacioeustatic drawdown; the distribution of neritic carbonates contracted almost to vanishing; (ii) the Blue Reflective in the Gippsland Basin (Figure 22) and its coeval equivalent DLS-5 on the northwestern Australian margin mark a sharp onset in developing incisions and canyons and downlapping reflectors; (iii) this is also the horizon of the last coals in the Latrobe coalfields and the top of the ‘coal-limestone sandwich’; and (iv) most generally this level marks the switch in conditions on the Australian continent from generally warmer and more humid to generally cooler and more arid.

CENOZOIC SEQUENCE IV: LATE Miocene TO HOLOCENE

Later Neogene stratigraphic and environmental succession in Australia

The Miocene marine record in southern Australia comprises a packet of sediments both culminating and terminating (in extent, biostratigraphic resolution, taxic richness of faunas and floras, and indicators of warming) in the middle Miocene Balcombian stage. Before ca 14 Ma, the record of strata is extensive and well dated. The record is not extensive again until the Early Pliocene. Within the intervening interval of ~8 million years we find some sediments in southeast Australia, and offshore in the southwest in ODP drilling. The same pattern can be seen in the western and northern margins, where there are bigger sheets of tropical carbonates and thick siliciclastics (Quilty 1977, 1980; McGowran 1979). In a striking and noteworthy correlation, there is a major change in New Guinea from carbonates to deepwater siliciclastics (McGowran 1979), shown by Haig and Mald (1996) to be Late Miocene in age in the Papuan Basin, while the shift between the two Neogene prograding wedges in the Gippsland Basin in the southeastern part of the continent is also Late Miocene (Figure 22). These patterns are more than eustatic: they have a strong regional tectonic component and match the two-part stratigraphic signature for the Neogene at large (Vail et al. 1991 figure 12), the low between two assemblages of third-order sequences falling within the Late Miocene.

The turnaround after the globally cool and regressive mid-Oligocene is well established by the broad pattern of the Miocene sea-level curve from the Late Oligocene to the Late Miocene (Figure 1), which has been called the Miocene oscillation (Frakes et al. 1987; McGowran & Li 1994, 1997a; McGowran et al. 1997a, b). The Leeuwin Current transported warm-water organisms from the New Guinea region to the southern margin, most strongly at the optimum, ca 16–15 Ma, thereby enhancing the warm signals of that time. This neritic pattern of maximum transgression across continental margins is bipolar, with a maximum spread by warm-water neritic biotas to higher latitudes at ca 16–15 Ma. The pattern is matched in the pelagic realm by migrations to high latitudes by sea-surface isotherms by up to 30° latitude. The warm peak is truncated by chill III, which is very sharp at ca 14 Ma in the Middle Miocene. The relatively cool regressive Late Miocene is well established. To the extent that we can correlate the faunas and floras of inland Australia with neritic and oceanic patterns, the rainforest-type floras and coal swamps and the rainforest-type faunas characterised Early Miocene time up to the climatic optimum, but then decline or disappear from the record (Martin 1991; Archer et al. 1994, 1995; Kershaw et al. 1994). The extensive warm-neritic and warm-terrestrial biotas contract and disappear simultaneously, both responding to the major global cooling at ca 14 Ma (chill III: Figure 9). Younger Miocene marine strata and events are known only from the southeast corner of the continent, and the terrestrial record loses resolution and quality.

Marine transgressions resume in the Early Pliocene record, and the climatic impact of this reversal of global cooling and marine regression was enhanced by the rejuvenated Leeuwin Current. There is a broad two-part pattern in the marine Pliocene, poorly constrained in the inner neritic (Lindsay 1981), but seen better in more open-marine facies (Gallagher et al. 2003). Hodell and Venz (1992) identified an interval of higher than average δ18O values at 4.1–3.9 Ma at Subantarctic ODP site 704, which correlates approximately with the sea-level lowstand terminating the Early Pliocene record (Haq et al. 1987).

Rainforests returned to southeastern Australia after a retreat during Late Miocene time (Martin 1991). The warm-wet rebound was marked and clear, but did not attain the

**Figure 24** A downcut reconstructed from three sections in the Latrobe Valley coalfields in Gippsland (Holdgate et al. 1995). This is a third-order cut; the Y1–2/Y3–4 boundary is shown and correlated in Figures 21, 23, 26; the colour index traces Milankovitch cycles (not yet studied cyclostratigraphically). Such downcuts can exceed 50 m in relief.
Figure 25 Regional and global Middle Miocene events: parallels between the three environmental realms. (Upper) Chronological parallels calibrated against Berggren et al. (1995) M, low-latitude and Mt, mid-latitude zones; Taylor, southeast Australia zones G–C; Li et al. (2003b, 2004) southern Australian SAN zones; and important and defining species events: 3rd-order SBs, sequence boundaries (Hardenbol et al. 1998); δ13C(+)ves, positive isotopic carbon excursions CMI–7 (Woodruff & Savin 1991) within the overall Monterey carbon excursion; δ18O(−)ves, negative isotopic oxygen excursions A–F (Woodruff & Savin 1991) interleaved with positive excursions labelled as third-order glacials Mi1b–Mi5 (Miller et al. 1998). Holbourne et al. (2004) identified these events off the western and southern margins. Two Lakes Entrance curves: a biofacies proxy for sea-level and a cancellate/spinose planktonic proxy for fertility (Li & McGowran 2000: scaled to depth not time). Large photosymbiotic neritic foraminifers: southern margin, McGowran (1979) and Lindsay (1985); northwestern margin, assemblages LF6–LF8, Chaproniere (1984). Incising and canyon-cutting, North West Shelf and southeastern Australia: in the two regions strong changes in seismic character occur at the same time, marked by reflector DLS-5 (Cathro et al. 2003; Moss et al. 2004) and the Blue Reflector (Holdgate et al. 2000a, 2003b: Figure 22), respectively. SE coalfields, coal sequences cease abruptly at the ‘top limestone–coal sandwich’ (Figure 26). Heavy line: onset of rapid sustained growth of West Antarctic icecap and shift from warmer more humid to cooler more arid conditions recorded in pelagic, neritic and terrestrial facies in both hemispheres. The change in seismic character in deep-neritic to bathyal facies is coeval with these more obviously exogenic phenomena. (Middle) Carbonate facies circum-Australia, showing the consistency of hiatuses on the three margins that can be correlated with third-order sequence boundaries; also the cutoff within the Serravallian. Thicknesses, ignored here, vary by an order of magnitude. (Lower) North American Land Mammal Ages (northern, terrestrial) and chronofaunas compared to Tethyan chronofaunas (neritic, photosymbiotic foraminifers) and New Zealand taxic cycles (southern and marine-benthic foraminifers). Boundaries are thick cables as a reminder that chronofaunal changeovers are not instantaneous. North American succession; Webb and Opdyke (1995), Woodburne and Swisher (1995), Prothero (1995), Janis et al. (2004). Deep-ocean benthic foraminiferal succession: Thomas et al. (2000). Changeover was prolonged and broadly coeval with Monterey carbon excursion (Figure 23). Event at ca 14 Ma across the environmental realms: McGowran and Li (2000, 2002).
levels of the Miocene in the Latrobe Valley (mean annual temperature 2–4°C above today’s, mean annual precipitation 50–70% above today’s; Gallagher et al. 2003). Tedford (1994) recorded the only Pliocene terrestrial vertebrate faunas in southeastern Australia hosting known rainforest taxa. As in the Miocene, the pattern is coherent, and a three-way equivalence in variation of sea-level, temperature and vegetation/precipitation still held. Thus Martin (1991 figure 5, 1994 figure 7.2) showed how the Late Miocene drop in precipitation and subsequent Early Pliocene increase (inferred from vegetation changes in the eastern Murray River catchment region) correspond to a sea-level fall (on the southern Australian margin) and subsequent rise, respectively, and to parallel fluctuations in sea-surface temperature. In the Late Pliocene widespread Araucarian forest and rainforest, with tropical taxa now extinct, contracted to be replaced by a modern-type mosaic of sclerophyllous forests, more open vegetation, and pockets of cool-temperate rainforest (Gallagher et al. 2003).

**Tectonic succession: Late Miocene basin inversion**

There is a major discontinuity above the Miocene carbonates in all southern basins as well as above the massive tropical carbonates along the northern margin. It was enhanced by a circum-Australian episode of crustal compression and basin inversion (Coblentz et al. 1985; Hill et al. 1995; Hills & Reynolds 2000). Cockshell (1995) and Perincek and Cockshell (1995) treated this as a Neogene phenomenon, but we can be chronologically more precise than that. The clearest manifestation of inversion where there are age constraints is between the early Middle Miocene and the Early Pliocene. Lindsay (1981, 1985) found that in regional epirogeny on the Para Fault in the St Vincent Basin the tilting, faulting and erosion post-dated the Bairnsdalian (late Middle Miocene) part of the Port Willunga Formation and was prior to the Croydon facies of the Kooyonga Formation of Pliocene age. The low-angle Miocene–Pliocene unconformity could be constrained biostratigraphically to ca 10–5 Ma, but no closer than that (McGowran & Li 1997b). The change in the crustal stress regime was from predominantly tensional to predominantly compressional (Tokarev et al. 1998), such change a likely cause for uplift, subsidence, or basin inversion (Sandiford 1999).

Dickinson et al. (2001, 2002) have reviewed the stratigraphic relationships in southeastern Australia and especially the age determinations on the unconformity/

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**Figure 26** Coals, regional transgressions and ice sheets. Coals cluster in some sequences but are absent from others, most notably from the Tortachilla at the first incursion by tropical foraminifers (*) and the Jan Juc at the onset of Oligocene warming (Late Oligocene warming event: LOWE). With those exceptions, coals correlate with benthic large foraminifers. Coeval coals and limestones cluster within sharp boundaries in the late Middle Eocene and Middle Miocene, hence the ‘limestone–coal sandwich’. The ice-volume curve: the δ18O signal minus an inferred temperature component based on the Mg/Ca ratio left an ice volume (or salinity signal) as residue (Lear et al. 2000; cautioning coarse resolution). This is controversial (Oerlemans 2004) but correlates with the four Cenozoic chills (Figure 2) and is in agreement with regional neritic and non-marine patterns. Coals are mainly from Holdgate et al. (1995, 2000b) and Holdgate and Clarke (2000). Transgressions and large foraminifers, mainly from McGowran et al. (1997a, 2000). Two grey spots with error bars as in Figure 2: clustered dates of two regolithic ferruginisations, implying episodic deep weathering (Pillans 2002).
hiatus, confirming with new foraminiferal and strontium data the ca 10–5 Ma gap but not refining it. Within this time there was deformation (gentle folding and reverse faulting with mostly 1–5° angularity with the overlying Pliocene), regional uplift in and around the sedimentary basins and removal of hundreds of metres of section, and exposing basement highs so that the Pliocene sediments are markedly richer in coarse siliciclastics than the Miocene carbonate and coal facies. A section through the Gippsland Basin onshore (Figure 27) is one of several that could be used to display the angular unconformity in the eastern half of southern Australia.

Late Miocene basin inversion is shown together with key events during the Cenozoic Era on the northern and southern continental margins (Figure 28).

Oceanic and intraplate deformation and plate-margin deformation

Drilling on ODP leg 116 in the Central Indian Ocean showed that a well-developed seismic unconformity of Late Miocene age could be dated at 8–7 Ma (Cochran 1990). This unconformity heralds the onset of an episode of widespread intraplate deformation. High-angle reverse faults have been traced from within oceanic basement (down to ~Moho depth) up into originally planar oceanic sediments, the faults being interpreted as a reactivation of two sets of pre-existing, spreading-centre, normal faults (Bull & Scrutton 1990; Bull 1990). The mode of deformation was shown to be a buckling of the entire brittle layer of the oceanic lithosphere (Bull et al. 1992).

Although it is clear that New Guinea includes several terranes accreted from the Melanesian Arc, the timings of the collisions are not so clear (Abbott et al. 1994, 1997; Crowhurst et al. 1996; Hill & Raza 1999). At the convergence of the western continental margin with the northern margin in the Timor Sea, the Australian Plate has been colliding with the Banda Arc, inducing left-lateral, oblique-slip faulting and the development of pullapart basins (the Ashmore–Cartier and Malita pullapart grabens). This was a geologically relatively sudden, Late Neogene phenomenon (Shuster et al. 1998) the synchrony of which, or otherwise, with events elsewhere on the plate can be tested by constraining the event itself. A change to oblique Australia–Pacific convergence at ca 12–10 Ma resulted in kilometres of uplift in northern New Guinea (Hill & Raza 1999). Haig and Medd (1996) used biofacies and biostратigraphy to describe Early Pliocene cycles in the Papuan Basin, shallowing upward and beginning at ca 6 Ma with a shallowing of more than 1000 m. They interpreted them as tectonic pulses, not eustatic cycles (although that was the time of the Messinian drawdown), responding to the onset of compressional deformation and uplift in the frontal part of the Papuan Fold Belt. Haig (1996) showed more generally in mainland Papua New Guinea that the pronounced tectonic and topographic changes of the Late Neogene, such as the collapse of neritic carbonate platforms to bathyal depths, commenced at ca 8 Ma. Timing was similar in New Zealand: compressive force became active at ca 12 Ma, plate convergence rapidly increased at ca 6–5 Ma and rapid uplift of the Southern Alps began (Walcott 1998).

The Late Miocene tectonic event bears heavily on structural risk in exploration on the Australian continent due to the formation of new, and reactivation of older, structural hydrocarbon traps with their modification and

Figure 27 Borehole correlations, Latrobe Valley and Gippsland Basin onshore (Dickinson et al. 2001, 2002), palynological and planktonic foraminiferal ages, Partridge (1971), Mallett (1977) and Dawson (1983). Late Miocene planktonic foraminiferal and palynological determinations are all pre-deformation and pre-unconformity; Pliocene palynological determinations are above and younger. This is the most detailed reconstruction available of the circumcontinental Miocene–Pliocene unconformity, which hosts a hiatus of ~10–5 million years. depn, depression; fmn, formation; mbr, member.
perhaps breaching. Baillie and Jacobson (1995) emphasised that this was one of the most important, economically speaking, of all the events in the Australian geological record.

Regional uplift in the late Neogene
The Flinders–Mt Lofty Ranges and their topographic contrast with flanking basins grew in the Late Neogene in the new, predominantly compressional regime, fixing upon a zone of thermally induced relative crustal weakness (Tokarev et al. 1998). The southeastern highlands have long been regarded as youthful (Plio-Pleistocene) and known as the Kosciuszko Uplift; however, extensive dating indicated a much earlier origin, leaving little understanding of Late Neogene uplift (Sandiford 2003). Using stratigraphy, Dickinson et al. (2001, 2002) could distinguish between two successive tectonic pulses, one before and one after the Early Pliocene transgression (labelled Jemmys Point on Figure 5), the earlier pulse being the more significant as to regional uplift, erosion and exhumation stripping hundreds of metres of section (thus according with the above comment on general significance in continent-wide petroleum geology). Using geomorphic analysis and the Early Pliocene transgression as a benchmark, Sandiford (2003) estimated some 175–240 m of uplift in the Otway Range since the Early Pliocene, most probably between 2 and 1 Ma.

Hiatuses, mass wasting, canyon cutting
Seismic stratigraphy indicates that incision or canyon cutting into carbonate-rich strata began at approximately the same time on opposite margins (North West Shelf and Gippsland Basin), that time being late Middle Miocene (Holdgate et al. 2000b; Gallagher et al. 2001; Cathro et al. 2003; Moss et al. 2004) at sequence boundary Ser2 (Figure 25). Li et al. (2004) identified three mega-hiatuses at ca 16–15 Ma, 9–8 Ma and 2.0–1.5 Ma. For the first of these they advanced arguments of differential uplift/subsidence in the Great Australian Bight (but not in southeast Australia). The second mega-hiatus falls within the time of tectonic events discussed above. The third also nestles among an array of Plio-Pleistocene tectonic and topographic events. Li et al. (2004) argued that the Great Australian Bight hiatuses recorded slope failure and slumping due to tectonic instability, itself a stepwise pattern due to pulsed shifts in ocean-floor spreading. It is likely that a better ordination of canyon cutting on Australian margins will reveal a pulsing there too.
Meanwhile, there is another climatic effect to consider: namely, that the biggest and most common cut-and-fills in southeast Australia are Late Neogene in age, Pliocene–Pleistocene, coeval with the sudden increase in carbonate accumulation rates on the margin and upper slope, common to all southern margins but demonstrated in the Great Australian Bight (Peary et al. 2000; Li et al. 2004). Prolific growth of bryozoans was due to high fertility in ice-age (lowstand) upwellings (James et al. 2004), and a thick, unstable and slump-prone sediment pile near the margin.

Earlier canyon cutting has been identified recently in the western Otway Basin. Leach and Wallace (2001) inferred an early Miocene onset of cutting in the Otway Basin. Further west where Early Oligocene carbonates are the best developed in southern Australia, Pollock et al. (2002) concluded that Early Oligocene carbonates were cut by canyons with fills of Late Oligocene age, the cut-and-fill continuing through the Neogene.

CONCLUDING REMARKS

Southern Australia is a trailing or passive continental margin displaying the mild, episodic diastrophism that those terms imply, but influenced by events to the north of Australia, the leading or active margin (Figure 28). Its terms imply, but influenced by events to the north of Australia. The Late Neogene sequence records the shaping of modern continental topography and the aridification of Australia. The Late Palaeogene sequence shows several responses to the 43 Ma global tectonic event: formation of rejuvenation of sedimentary basins, the accumulation of neritic carbonates and voluminous paralic coals, a tectonic signature on several third-order hiatuses, oceanic influence as the Southern Ocean widens more rapidly, and warming, the last two impacting on the biotas. The Early Neogene sequence marks the apogee of marine transgression and the second act in the association of coals and neritic carbonates. This was the time of significant collisions in the north, with far-reaching effects on palaeoceanography, climate and biogeography. The Late Neogene sequence records the shaping of modern continental topography and the aridification of Australia. Pervading general statements about the stratigraphic sequences are three overriding generalisations: southern Australia’s physical geology boils down to its episodic, northwards continental drift; its environmental history has a strongly global stamp; and its biotic history is a mixture of both.

The outstanding stratigraphic problems include the following: (i) gentle deformation means poor outcrop and very patchy subsurface sampling both onshore and offshore; (ii) biostratigraphy in the neritic realm is insufficiently integrated with magnetostratigraphy and strontium isotopic stratigraphy; and (iii) cross-correlations between the neritic realm and the vertebrate faunas and macrofloras of the continental realm are all too tenuous.

However, there are excellent reasons for addressing these problems in the near future. One is that the southern margin is at the sensitive mid-latitudes facing the newly widening Southern Ocean, the engine-room of the cooling planet, with implications for the biotas and fossil record in all three environmental realms as well as the history of the regolith. A second reason is sequence-stratigraphic: to test the relative influence of tectonism and eustasy on the widespread unconformities and transgressions. A third reason is to develop our notions of the where and when of the coal–limestone third-order couplets making up the coal–limestone sandwich. Progress in correlation and age determination, both with global scales and, more urgently, between the environmental realms, should focus numerous other problems in sedimentology, palaeobiology and regolith geology.

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