The Tasmanides of eastern Australia

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Abstract: The Tasmanides of eastern Australia record the break-up of Rodinia, followed by the growth of orogenic belts along the eastern margin of Gondwana. Spatially, the Tasmanides comprise five orogenic belts, with an internal Permian–Triassic rift–foreland basin system. Temporally, the Tasmanides comprise three (super)cycles, each encompassing relatively long periods of sedimentation and igneous activity, terminated by short deformational events. The Neoproterozoic–earliest Ordovician Delamerian cycle began by rifting, followed by convergent margin tectonism and accretion of island-arc forearc crust and ?island arcs in the Middle–Late Cambrian. The Ordovician–Carboniferous convergent margin Lachlan supercycle consists of three separate cycles, each ending in major deformation. The Ordovician Benambran cycle includes convergent (island-arc) and transform margin activity terminated by terrane accretion in the latest Ordovician–earliest Silurian. The Silurian–Middle Devonian Tabberabberan cycle reflects development of a large back-arc basin system, marked by rift basins and granite batholiths, behind intra-oceanic arcs and an Ordovician–Early Devonian terrane that were accreted in the Middle Devonian. The Middle Devonian to Carboniferous Kanimblan cycle began by rifting, followed by continental sedimentation inboard of a major convergent margin system that forms the early part of the Late Devonian–Triassic Hunter–Bowen supercycle. This supercycle comprises a Late Devonian–Carboniferous continental arc, forearc basin and outboard accreted terranes and subduction complexes intruded by the roots of a Permian–Triassic continental margin arc. Complex deformation ended with accretion of an intra-oceanic arc in the Early Triassic. Key features of the Tasmanides are: continuity of cycles across and along its length, precluding growth by simple eastwards accretion; development of a segmented plate margin in the Late Cambrian, reflected by major rollback of the proto-Pacific plate opposite the southern part of the Tasmanides; rifting of parts of the Delamerian margin oceanwards, to form substrate to outboard parts of the Tasmanides; the presence of five major Ordovician terranes in the Lachlan Orogen; and the generation of deformations either by the accretion of arcs, the largely orogen-parallel 'transpressive' accretion of Ordovician turbidite terranes (in the Lachlan Orogen), or by changes in plate coupling.

The Tasmanides of eastern Australian represent one sector of the Pacific margin of Gondwana that stretched 20 000 km through New Zealand, Antarctica (North Victoria Land, Transantarctic Mountains, Antarctic Peninsula) and into South America. This paper reviews that history and discusses tectonic interpretations of the Tasmanides in three parts: Part 1 introduces the concept of the Tasmanides; Part 2 presents an up-to-date synthesis of its development; Part 3 focuses on some key issues and processes distilled from that synthesis. This paper draws on published, or in press, papers on specific aspects of the Tasmanides, as well as building on previous syntheses in Coney (1992), Scheibner & Veevers (2000), and Betts *et al.* (2002), and of Scheibner & Basden (1998) for New South Wales, VandenBerg *et al.* (2000) for Victoria, Gray & Foster (1997) largely for Victoria, Seymour & Calver (1995) for Tasmania, Bain & Draper (1997) for North Queensland and Crawford *et al.* (2003*a*, *b*) for the Cambrian of Tasmania and Victoria.

Part 1: The Tasmanides – definitions

Tasmanides is the name given to a collection of orogenic belts that records the break-up of a Mesoproterozoic supercontinent, the formation of a passive margin in the Late Neoproterozoic and the establishment of a series of convergent margin orogenic belts along part of east Gondwana from the Middle Cambrian, through collision of Gondwana with Laurussia to form Pangaea *c.* 320–330 Ma ago (Veevers 2000*b*), until the beginning of Gondwana–Pangaea break-up, around 227 Ma.

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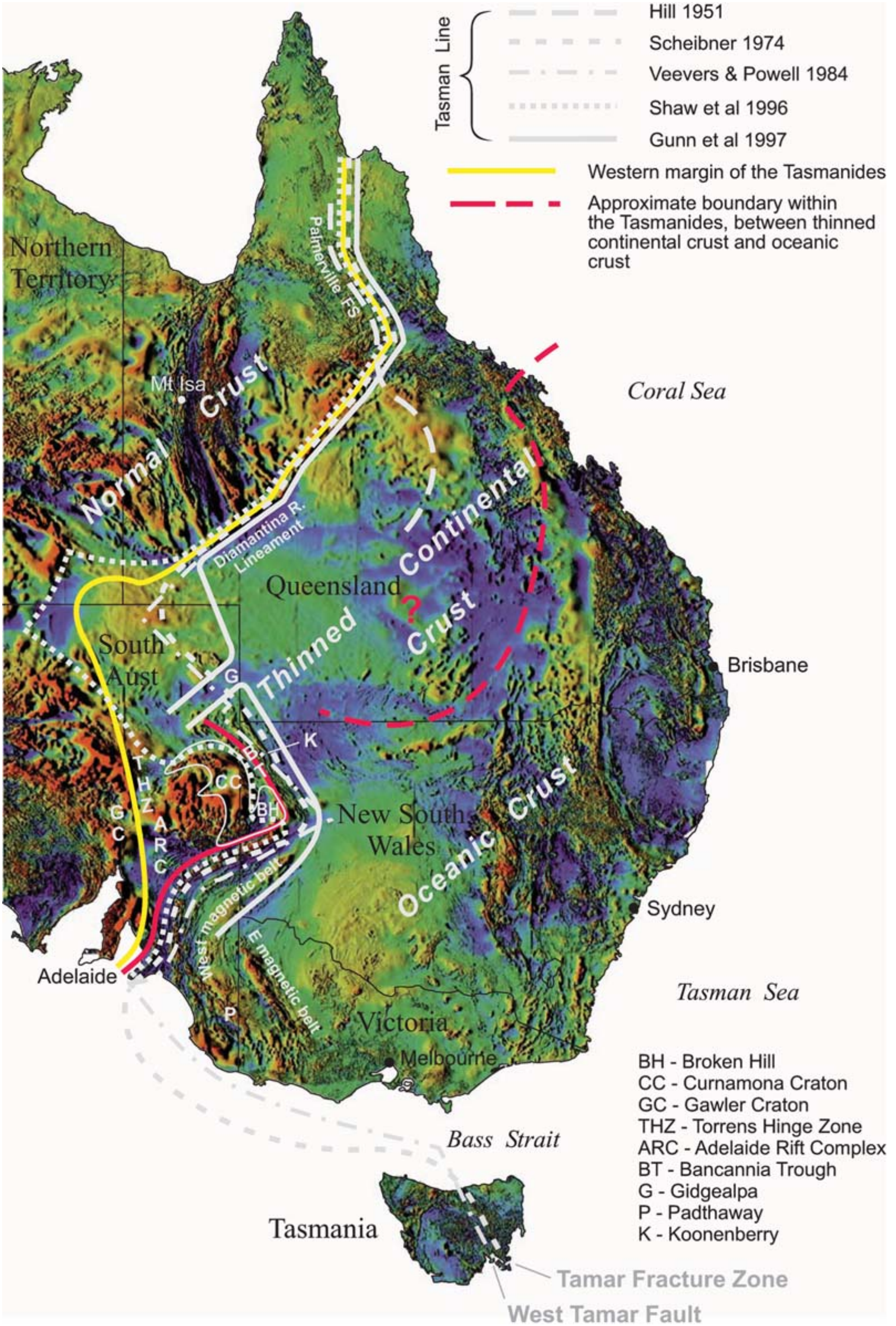

Tasman Line
Hill 1951
Scheibner 1974
Veevers & Powell 1984
Shaw et al 1996
Gunn et al 1997
Western margin of the Tasmanides
Approximate boundary within the Tasmanides, between thinned continental crust and oceanic crust
Northern Territory
Mt Isa
Normal Crust
Palmerville FS
Coral Sea
Diamantina R. Lineament
Queensland
Thinned Continental Crust
South Aust
G
K
CC
BH
THZ
GC
ARC
New South Wales
Oceanic Crust
Brisbane
Sydney
West magnetic belt
E magnetic belt
Adelaide
P
Victoria
Melbourne
Tasman Sea
Bass Strait
Tasmania
Tamar Fracture Zone
West Tamar Fault
BH - Broken Hill
CC - Curnamona Craton
GC - Gawler Craton
THZ - Torrens Hinge Zone
ARC - Adelaide Rift Complex
BT - Bancannia Trough
G - Gidgealpa
P - Padthaway
K - Koonenberry

Extent of the Tasmanides

Where and what are the boundaries of the Tasmanides? The eastern boundary was rifted off by opening of the Tasman Sea, beginning at *c.* 90 Ma, and of the Coral Sea to the north at *c.* 61 Ma (Sdrolias *et al.* 2003). As a result, fragments of the Tasmanides are preserved in parts of New Zealand, New Caledonia and, presumably, the Lord Howe Rise (Sutherland 1999). The northern boundary lies in eastern and northern Papua New Guinea (Hill & Hall 2003). The western boundary is more difficult to define and, here, it is suggested that it is a structural boundary that represents both Proterozoic rifting and, thus, major thinning of Precambrian continental crust from a normal thickness of 38–41 km (Collins *et al.* 2003), as well as later extensional/reverse/strike-slip reactivation. The Torrens Hinge Zone along the western edge of the Adelaide Rift Complex represents a part of this margin that has undergone only minor contractional reactivation. In contrast, the Palmerville Fault System in northern Queensland represents part of the margin that has undergone major contractional reactivation in the Devonian–Carboniferous.

Using this definition, it is suggested that the western edge of the Tasmanides is a zig-zag line that crosses the Australian continent from north to south (Fig. 1). In northern Queensland, the western margin of the Tasmanides coincides with the N-trending Palmerville Fault System (Fig. 1). Shaw *et al.* (1987) showed that this fault system is a major west-dipping Devonian–Carboniferous thrust system with unknown amounts of displacement. In central Queensland, the western margin is drawn along the NE-trending Diamantina River Lineament. Although this lineament truncates the southern part of the Palaeoproterozoic to Mesoproterozoic Mt Isa succession, it has no surface expression. Further south, it is argued that the western line of the Tasmanides runs along the N–S Torrens Hinge Zone that separates the Gawler craton – with thin Neoproterozoic cover – in the west, from the 4 km thick Adelaide Geosyncline (e.g. Preiss 2000, herein called the Adelaide Rift Complex) to the east. This line runs N–S, west of Mesoproterozoic rocks of the Curnamona craton, which is bounded to the east and west by the Adelaide Rift Complex. The western margin of the Tasmanides disappears offshore south of the Australian mainland, with Tasmania lying east of that boundary. As thus defined, the western margin of the Tasmanides separates continental crust of 'average thickness' in the west from crust in the east that underwent thinning during supercontinent break-up, mainly in the Neoproterozoic.

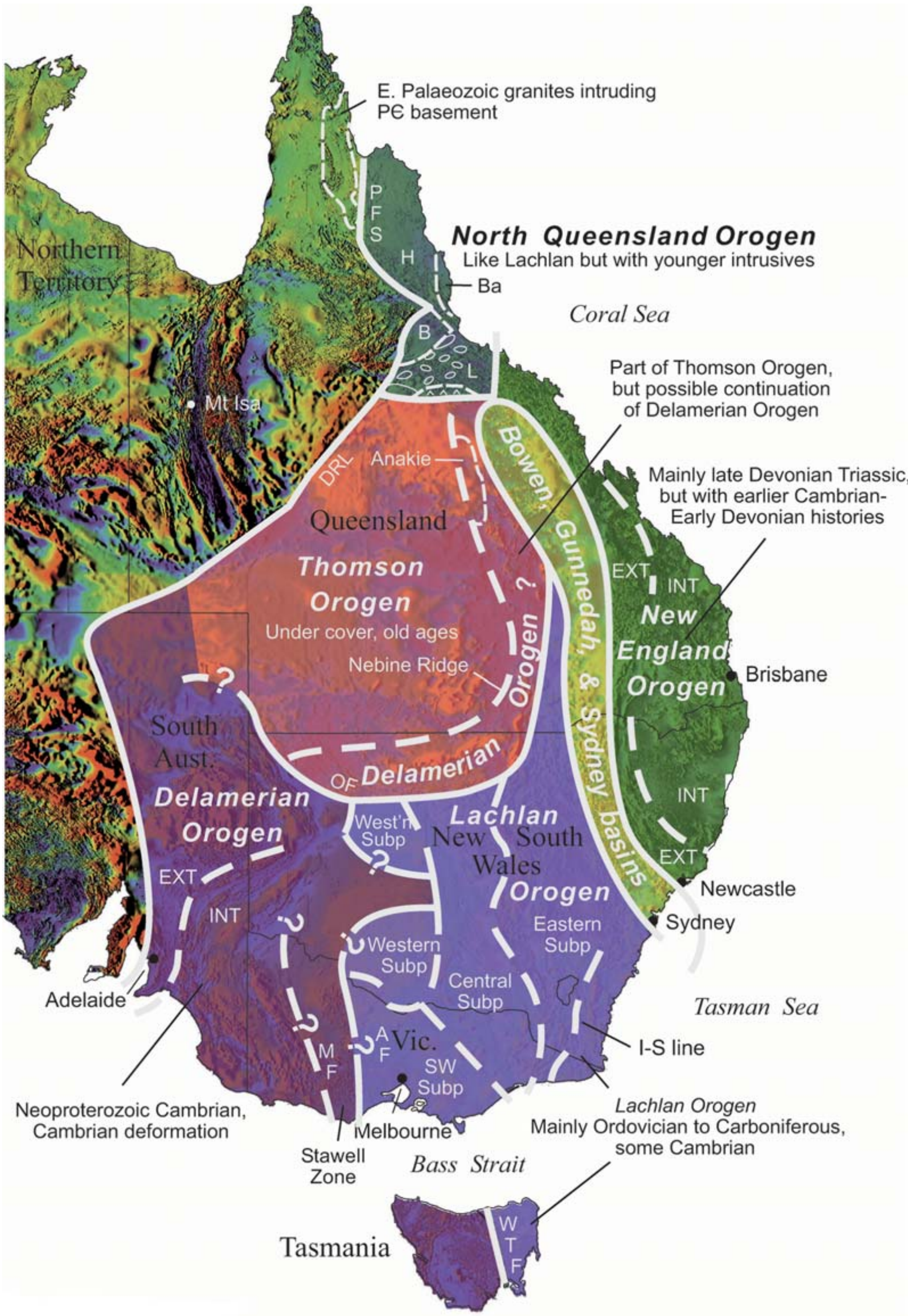
Except in Queensland, this boundary of the Tasmanides does not coincide with the Tasman Line, described originally by Hill (1951) as the eastern boundary of Precambrian 'basement' rocks in eastern Australia. Subsequently, the concept of the Tasman Line has been broadened to represent the western boundary of the Tasmanides, with the line considered to mark the place of break-up of a Mesoproterozoic supercontinent of which Australia was part (e.g. Veevers & Powell 1984; Powell *et al.* 1994; Scheibner & Basden 1998; Scheibner & Veevers 2000; but not Direen & Crawford 2003*b*). This paper does not subscribe to this view, as discussed in Part 3.

'Continent–ocean boundary' within the Tasmanides

Outboard of the western edge of the Tasmanides, it is possible to recognize a second line that marks the approximate boundary between extended continental crust and oceanic igneous crust (Fig. 1). This line runs along the northern and western edge of the Kanmantoo Trough and under the Bancannia Trough. Extension further north is uncertain. In far north Queensland, it lies outboard of the present-day coastline, since continental crust underlies the North Queensland Orogen (Bain & Draper 1997; Hutton *et al.* 1998). Offshore plateaux adjacent to the Queensland coast are composed of granitic intrusions into Devonian metasediments similar to that of the North Queensland Orogen (Feary *et al.* 1993). Between the Bancannia Trough and far north Queensland, the location of the 'continent–ocean boundary' is based on the inference that probable thinned continental crust underlies most of the Thomson Orogen (Finlayson 1990). There is general agreement that the accretionary parts of the New England Orogen developed on oceanic crust. The forearc basin of that orogen formed on arcs of the underlying Lachlan Orogen that had become accreted by the Middle Devonian.

In this interpretation, the Mesoproterozoic rocks of the Curnamona craton (encompassing

Fig. 1. Map of eastern Australia showing the Tasmanides, different locations of the Tasman Line (based on Direen & Crawford 2003*b*) and the 'continent–ocean boundary'. Geophysical image from Geoscience Australia (Milligan & Franklin 2004).



E. Palaeozoic granites intruding PЄ basement
P F S
H
North Queensland Orogen
Like Lachlan but with younger intrusives
Ba
Coral Sea
B
L
Northern Territory
Mt Isa
Part of Thomson Orogen, but possible continuation of Delamerian Orogen
Anakie
DRL
Bowen, Gunnedah, & Sydney basins
Mainly late Devonian Triassic, but with earlier Cambrian-Early Devonian histories
Queensland
Thomson Orogen
Under cover, old ages
Nebine Ridge
Orogen ?
EXT
INT
New England Orogen
Brisbane
South Aust.
OF Delamerian
Delamerian Orogen
West'n Subp
Lachlan
New South Wales
Orogen
EXT
INT
Newcastle
Sydney
Eastern Subp
Western Subp
Central Subp
Adelaide
Tasman Sea
A F
Vic.
I-S line
M F
SW Subp
Neoproterozoic Cambrian, Cambrian deformation
Melbourne
Lachlan Orogen
Mainly Ordovician to Carboniferous, some Cambrian
Stawell Zone
Bass Strait
Tasmania
W T F

the Mt Painter, Olary and Broken Hill areas) are surrounded by the Adelaide Rift Complex. Rutland (1976) suggested that the Adelaide Rift Complex was a multi-branched rift, with a central unextended basement horst, the Curnamona craton. If not, then one is forced to look at core-complex type extensional models to exhume the craton in the middle of a single rift system.

Subdivision of the Tasmanides

Rutland (1976) suggested that the Tasmanides constituted one of the major orogenic provinces of Australia (his Tasman Province), which was divided into a number of subprovinces: Delamerian, Lachlan and New England. Subsequent usage has been to elevate these subprovinces to province status, calling them orogenic belts (Fig. 2). Subdivisions of these orogenic belts are called subprovinces and, in some cases, these have been further divided into structural zones and blocks (e.g. in New South Wales by Owen & Wyborn (1979), Glen (1992) and Scheibner & Basden (1996) and in Victoria by Gray (1988) and VandenBerg *et al.* (2000)). Orogenic belts that constitute the Tasmanides are now described from west to east.

The Delamerian Orogen

The extent of the Delamerian Orogen is defined by the distribution of rocks that have undergone a multistage Mid–Late Cambrian to earliest Ordovician Delamerian deformation. The Delamerian Orogen thus encompasses western Tasmania, western NSW and Victoria, and eastern South Australia (Fig. 2). In western Tasmania, the Delamerian (or Tyennan) Orogen extends west from the West Tamar Fault zone of Reed *et al.* (2002) (Fig. 2) and contains rocks that range in age from Neoproterozoic to Early Ordovician, the latter being post-tectonic to the Tyennan Orogeny, which is equivalent in time and dynamics to the Delamerian Orogeny on the mainland.

On the mainland, the Delamerian Orogen extends westwards from western New South Wales and Victoria into South Australia, with the northern and eastern parts obscured largely by Mesozoic and Cenozoic cover. In the Koonenberry area, the eastern boundary is taken to be the Olepoloko Fault System (Stevens 1991) (Fig. 2). In western Victoria, the boundary has been controversial and has been moved over 200 km east and west over the last ten years (see VandenBerg *et al.* 2000). Although most authors had settled on an east-dipping thrust – the Moyston Fault (Fig. 2) (VandenBerg *et al.* 2000) – as the Delamerian/Lachlan boundary, 500 Ma (Delamerian) mica cooling ages in the Stawell Zone (Miller *et al.* 2003; 2004) suggest that the boundary may be the west-dipping Avoca Fault (Fig. 2). The whole of the Stawell Zone is either in the Delamerian Orogen [assumed in this paper, as suggested by Glen (1992) and Glen *et al.* (1992)], or in a transitional intermediate zone (Miller *et al.* 2004). The Stawell Zone was reactivated in the Ordovician–Devonian during the Lachlan supercycle. The eastern edge of the Delamerian Orogen in NSW is uncertain, being obscured by Mesozoic and Cenozoic cover. A recent interpretation (Hallett *et al.* 2005), suggests that the Delamerian Orogen extends northeastwards from Victoria with a cusp shape into western New South Wales and, in the north, lies partly as basement to the western part of the Lachlan Orogen (Figs 2, 6).

On the mainland, the Delamerian Orogen is divided into two parts (Fig. 2). In the west is an external fold–thrust belt, called the Adelaide Fold–Thrust Belt by some, that developed from inversion of the Adelaide Rift Complex and overlying shallow-water Cambrian sediments. Further east, the internal part of the Delamerian Orogen, called the Kanmantoo Fold Belt by some workers (e.g. Scheibner 1987), consists of multiply-deformed and metamorphosed sediments (intruded by granitoids) that developed from the inversion of the deep-water Cambrian Kanmantoo Trough. These metamorphic rocks and granitic rocks extend eastwards under Cenozoic cover into western New South Wales and Victoria. They pass eastwards into two elongate (350 × 50 km) belts of volcanic rocks

Fig. 2. Subdivisions of the Tasmanides. Also shown is the I–S line separating I-Type granitoids on the east from S±I-type granitoids on the west, subprovinces (subp) within the Lachlan Orogen (eastern, central, western and southwestern), subdivision of the Delamerian and New England orogens into internal (INT) and external (EXT) parts, and subdivision of the North Queensland Orogen into subprovinces (H, Hodgkinson; B, Broken River; L, Lolworth–Ravenswood: Ba, Barnard Metamorphics). DRL, Diamantina River Lineament; AF, Avoca Fault; MF, Moyston Fault; PFS, Palmerville Fault System; OF, Olepoloko Fault; WTF, West Tamar Fault. The Lolworth–Ravenswood block also shows basement inliers and Cambro-Ordovician volcanic rocks (^) of the Seventy Mile Group. Polygon west of the I–S line is the Australian Capital Territory.

concealed beneath Mesozoic and younger cover, as revealed by aeromagnetic data (Figs 1, 5a).

Basement to the Delamerian Orogen on the mainland is Palaeoproterozoic and Mesoproterozoic crust brought to the surface in the hangingwalls of some thrusts south of Adelaide. These rocks are inferred to be part of the Curnamona craton, exposed in the Broken Hill–Olary region (located in Fig. 1). Granite data (Foden *et al.* 2002) suggest that the Kanmantoo Trough (Fig. 5a) was developed directly on oceanic crust. Similarly, the presence of Proterozoic E-MORB basalts east of the Devonian Bancannia Trough in the Koonenberry area in the northeast of the orogen (Mills 2003) suggests that Cambrian turbidites there were developed on pre-break-up extensional volcanic rocks (A. Crawford pers. comm. 2004). Os isotopic data suggest that Proterozoic lithospheric mantle extends east into the Stawell Zone (Handler *et al.* 1997).

The Delamerian Orogen in Tasmania and on the mainland has undergone Carboniferous deformations, as well as Early and Middle Devonian heating, that reflect events in the neighbouring Lachlan Orogen to the east.

Lachlan Orogen

The Lachlan Orogen contains mainly rocks that range in age from basal Ordovician through to Carboniferous. It does, however, contain several belts of Cambrian greenstone in central Victoria that overlap in age with the Delamerian Orogen, as well as Cambrian rocks on the south coast of New South Wales. Deformation events occur around the Ordovician–Silurian boundary (Benambran Orogeny), locally around the Silurian–Devonian boundary (Bindi deformation), late Early to Middle Devonian (Tabberabberan Orogeny) and Early Carboniferous (Kanimblan Orogeny).

The orogen is divided structurally into four subprovinces (Figs 2, 6), separated by major faults. The boundaries between the Eastern and Central, and between the Central and Southwestern subprovinces are major oblique overthrusts and are called sutures because of disappearance of small amounts (hundreds of kilometres) of oceanic crust (e.g. Scheibner 1987). Although Gray and coworkers (e.g. Gray *et al.* 1997) do not distinguish between the Southwestern and Western subprovinces, interpretation of geophysical data indicates north-to-south differences in geology (Fig. 6).

A variety of possibilities has been put forward for the substrate of the Lachlan Orogen: continental (Rutland 1976; Chappell & White 1974), oceanic (Crook (1980), mixed oceanic and continental (Scheibner 1973), and oceanic in central Victoria with subsequent major underthrusting of continental material (Crawford *et al.* 1984). Traditional models of granite genesis suggested that west of the I–S line, the Lachlan Orogen was underlain by Precambrian continental crust (Chappell & White 1974), but results of detrital zircon dating indicate that this crust is no older than Early Ordovician (Williams & Chappell 1998). I-type granites east of the I–S line were thought to be sourced from 500–600 Ma tonalites (Chappell & Stephens 1988; Williams & Chappell 1998). Alternative models of granite genesis suggest that a mafic igneous substrate underlies Ordovician turbidites (Keay *et al.* 1997; Collins 1998) and this is consistent with high positive epsilon Nd data from the Ordovician Macquarie Arc (Glen *et al.* 1998; Crawford *et al.* 2005). In the Southwestern and Central subprovinces, conformable relations between Cambrian volcanic and sedimentary rocks and overlying Ordovician turbidites are used to infer that the turbidites were deposited on oceanic igneous substrate (Crawford *et al.* 1984; Gray & Willman 1991*a*).

Some authors have suggested that rifted-off fragments of Precambrian craton form the substrate to parts of the Lachlan Orogen. Scheibner (1989) suggested that there was a 'Molong microcontinent' under the Ordovician Macquarie Arc because of the shoshonitic affinities of some of the volcanism (based on the interpretation of Wyborn 1992). However, shoshonitic volcanism, now known to be restricted largely to the Late Ordovician in that arc (Glen *et al.* 1998; Crawford *et al.* 2005), is also present in modern intra-oceanic volcanic arcs such as Fiji. A continental substrate is precluded effectively by the high positive epsilon Nd values (Glen *et al.* 1998; Crawford *et al.* 2005) and primitive Pb isotope ratios (Carr *et al.* 1995). Packham (1973) and Scheibner (1989) also suggested that there was Precambrian continental crust under central Victoria: the latter basing his interpretation on the view that the thin-skinned thrusts described from the western, Bendigo Zone of the Southwestern subprovince (Figs 2, 6) by Cox *et al.* (1991) were foreland-type thrusts developed on rigid basement. Although Gray & Willman (1991*a*, *b*) and Gray (1995) showed that these thrusts were not of a foreland type (see also Glen & VandenBerg 1996) and could develop on oceanic igneous crust, Cayley *et al.* (2002) resurrected the idea of a continental basement (Selywn

Block), based on apparently continuous magnetic zones from western Tasmania towards central Victoria.

The Thomson Orogen

The Thomson Orogen underlies much of central and western Queensland where it is concealed by Mesozoic cover. The western margin against obscured rocks of the Delamerian Orogen is uncertain (Murray 1994): both contain low-grade rocks. As suggested from geophysical trends by Murray & Kirkegaard (1978) and Wellman (1995), the southern boundary of the Thomson Orogen against the Lachlan Orogen is a curvilinear east–west fault zone in north-western New South Wales (Fig. 2), (Olepoloko Fault in the west, Louth-Eumarra Shear Zone in the east: Stevens 1985 and Glen *et al.* 1996, respectively). This boundary truncates N–S structures of the Lachlan Orogen at a high angle, and curved and concealed magnetic volcanic units in the southern Thomson Orogen at a low angle (Fig. 2). This fault zone is inferred to be a major suture between two orogenic belts (see below). It swings northwards into Queensland where it may correspond to the Foyleview Geosuture of Finlayson *et al.* (1990). Further north, the eastern boundary of the Thomson Orogen extends east to the Permian–Triassic Bowen Basin on the surface. Interpretation of seismic reflection data (Korsch *et al.* 1997) suggests that the orogen passes beneath that basin, to underlie structurally the western part of the New England Orogen. To the north, the Thomson Orogen is bounded by the North Queensland Orogen and Meso- and Palaeoproterozoic cratonic Australia.

Drill hole data indicate that rocks in the Thomson Orogen range in age from Precambrian through to Late Devonian (Murray 1994; Scheibner & Veevers 2000). However, Cambrian or older rocks deformed in the 500 Ma Delamerian Orogeny occur along the eastern margin, in the Anakie Inlier (Fig. 2). This inlier extends in the subsurface to the south, into the Nebine gravity ridge (Withnall 1995). Cores from this ridge consist of metasediments, phyllites and multiply-deformed schist at lower greenschist to ?amphibolite grade (Murray 1994). However, the ages of these rocks are unknown, the only indicator being a 416 Ma K–Ar date on biotite (Murray 1994).

The south end of the Nebine ridge swings to the west and coincides with belts of major aeromagnetic and gravity highs that reflect concealed igneous rocks (proven by drilling, but of unknown age) that lie along the southern boundary of the orogen (Fig. 2, see below). It is suggested here that they may be of Cambrian age, since there is some evidence (below) that the oldest deformation across the boundary between the Lachlan and Thomson orogens is Late Cambrian, probably reflecting accretion of an intra-oceanic arc. When coupled with the uncertain nature of the western boundary of the Thomson Orogen (above), this continuity allows two possible suggestions.

1. The Nebine ridge and the Anakie Inlier are physically part of an enlarged Delamerian Orogen that extended eastwards for some 600 km. Such a margin is similar to the Nebine arc inferred by Harrington (1974). If validated by future work, such an orientation implies that a major sector of the Australian plate margin was orientated east–west in the Cambrian.
2. Older rocks along the southern and eastern margin of the Thomson Orogen are pieces rifted away from Delamerian Gondwana, with the Thomson Orogen being the site of post-500 Ma rifting. In this model, the Thomson Orogen is floored by extended and thinned continental crust (Harrington 1974; Murray 1994) that might correspond to the geophysically layered crust of Finlayson *et al.* (1990). Gravity data show that this crust has WNW trends in the southern third of the orogen and NE trends in the north (Milligan *et al.* 2003).

The New England Orogen

The New England Orogen is the most easterly component of the Tasmanides. It occupies much of coastal Queensland and extends south below the Mesozoic cover of the Clarence–Moreton and Surat basins into northeastern New South Wales (southern New England Orogen) (Figs 2, 13a; Leitch 1974; Scheibner & Veevers 2000). The orogen also forms basement to the eastern part of the Sydney Basin and extends offshore as the Currarong Orogen (Jones & McDonnell 1981; Jones *et al.* 1984), which is represented in seismic data as the ‘offshore uplift’ (Bradley 1993; Alder *et al.* 1998). The New England Orogen has an inferred thrust contact with the Eastern subprovince of the Lachlan Orogen.

The New England Orogen is divided into two structural subprovinces (Leitch 1974): a western, external part that constitutes a fold–thrust belt (with intrusive granites in the north) and an internal part in which accretionary complex rocks are multiply-deformed and metamorphosed, and intruded by granitoids. This subdivision reflects the development of a Late Devonian–Carboniferous classical convergent margin consisting of arc, forearc basin and accreted terranes. The subsequent history involves Permian rifting and Triassic subduction followed by a protracted Permian–Triassic (Hunter–Bowen) deformation.

Substrate to the New England Orogen is inferred to be oceanic, east of the Peel–Manning Fault System in New South Wales and the Yarrol Fault System in Queensland, and mixed oceanic and continental to the west, becoming continental by the Carboniferous and undergoing rifting in the Early Permian. However, small fault-bounded outcrops of Neoproterozoic–Early Devonian subduction-related and accretionary material overlap with the development of Delamerian and Lachlan orogens and point to the presence of older substrate. This is consistent with several other lines of evidence:

1. presence of Neoproterozoic Re–Os ages, commonly in the range of 0.6 to.1.8 Ga (Powell & O'Reilly 2001; Bennett *et al.* 2002);
2. Neoproterozoic zircon model ages for some young granites (Shaw & Flood 2002);
3. seismically fast lithospheric mantle interpreted from SKIPPY data by Van der Hilst *et al.* (1998); and
4. *c.* 562 Ma Sm–Nd isochron age on gabbro in ophiolitic rocks of a mid-Palaeozoic accretionary complex in central Queensland (Bruce *et al.* 2000).

Together, they point to the persistence of old lithosphere.

Permian–Triassic Bowen–Gunnedah–Sydney Basin system

This system is remarkable for its length of *c.* 1600 km (Fig. 2). It originated by rifting in the Early Permian and was converted into a basin foreland in the mid-Triassic, with the eastern parts constituting the foreland fold–thrust belt of the New England Orogen (Glen & Beckett 1997). The geology has been summarized by several authors (Harrington *et al.* 1989; Veevers *et al.* 1994; updated in Veevers 2000*d*, Murray 1990; Scheibner & Basden 1998; Fielding *et al.* 2001).

The substrate of the Bowen Basin is the Thomson Orogen in the west and New England Orogen in the east. The Gunnedah Basin was built over crust of the Lachlan Orogen (Korsch *et al.* 2002). The substrate of the Sydney Basin is inferred to be Lachlan Orogen in the west (felsic volcaniclastic rocks have been brought up in diatremes that intrude the basin, O'Reilly 1990) and the New England Orogen in the east, with a depositional relationship preserved north of Newcastle (Roberts & Engel 1987).

The North Queensland Orogen

This term has been introduced in order to group together a series of tectonic provinces (which have been downgraded to subprovince status) in far north Queensland, known as Hodgkinson Province, Broken River Province, Lolworth–Ravenswood Block and Barnard Province (Fig. 2). These were reviewed most recently by Bain & Draper (1997). The broad Ordovician to Carboniferous evolution of the North Queensland Orogen is similar broadly to that of the Lachlan Orogen, with two important exceptions – the presence of scattered inliers of Neoproterozoic continental crust that reflect growth on Precambrian continental crust, and the presence of Cambrian to Ordovician arc-related rocks in the south that are similar to the Tasmanian part of the Delamerian Orogen.

To the west, the North Queensland Orogen is fault-bounded against Meso-and Palaeoproterozoic rocks of cratonic Australia along the Palmerville Fault System, an imbricate Devonian–Carboniferous thrust system (Shaw *et al.* 1987). The North Queensland Orogen extends offshore, as indicated by inferred Palaeozoic strata underlying the offshore Marion and Queensland plateaux (Feary *et al.* 1993), although Henderson *et al.* (1998) suggested that these plateaux were underlain by accretionary complex rocks of the New England Orogen. The southern boundary against the Thomson Orogen is marked by truncation of the N-trending pre-500 Ma Anakie Inlier, seen also in gravity images (Murray *et al.* 1989; Milligan *et al.* 2003), although similar-aged Precambrian crust occurs as inliers within the orogen. The North Queensland Orogen underwent deformation in the Ordovician, Middle Devonian and Carboniferous (Bain & Draper 1997).

Tectonic cycles

The subdivision into different orogenic belts is based on the age distribution of constituent strata and the ages of 'climactic' deformation. The eastwards-younging of ages in those belts (Fig. 2) has been cited widely as evidence that the Australian craton grew by eastward accretion from the Neoproterozoic until the plate boundary jumped to the New Caledonia–New Zealand area in the Mesozoic (Sutherland 1999).

However, the increasing recognition of older rocks in parts of the so-called younger orogenic belts shows that growth of the Australian sector of Gondwana was not so simple. Cambrian rocks occur in parts of the Lachlan and New England orogens. They also occur, or are inferred, from the eastern and southern parts of the Thomson Orogen. This means that subdivision of the Tasmanides into orogenic belts based solely on age criteria is not as valid as thought previously. While it is still useful to show the different orogenic belts on a map (Fig. 2), time–space plots show old rocks extend right across the Tasmanides (Fig. 4).

It is thus proposed here to describe the evolution of the Tasmanides in terms of tectonic cycles, in the South American sense, to encompass depositional/magmatic as well as contractional deformational histories of rock packages developed along plate margins. A complete idealized cycle begins with a rift phase that may pass into drift, then into convergence (if subduction can be recognized), then into deformation/collision that terminates the cycle although a post-collisional phase may be present. In some cases there is no rift/drift phase. The term ‘collision’ is applied to the terminal deformation of many cycles in the belief that it reflects the accretion of an arc to the developing Gondwana landmass. It does not represent a continent–continent collision.

Acenolaza & Toselli (1981) and Ramos (1999) have described the cycle concept for the Palaeozoic of the South American margin of Gondwana. Seven out of the eight cycles there close with a deformation event. The exception is the Famatinian cycle that includes two significant deformations in the Ordovician (at 465 Ma and *c.* 440 Ma). The use of cycles here is similar to the concept of stages used previously in New South Wales (Scheibner 1973; Scheibner & Basden 1998) and in the New England Orogen by Korsch & Harrington (1981).

Part 2: Synthesis of the Tasmanides

This synthesis is based largely on new datasets collected in the last ten years, in the period since the *Tectonophysics* issue on the Lachlan Fold Belt and related regions (Fergusson & Glen 1992). Key datasets include new aeromagnetic and gravity databases, new geological mapping, new mineral and *in situ* age-dating techniques, new conodont identification techniques by Ian Stewart (Monash University) and the collection of deep seismic reflection profiles.

Evolution of the Tasmanides is now described from the oldest cycle to the youngest, proceeding from south to north and west to east. Each cycle is described first from its ‘home’ orogen and then from other orogens (such as the Delamerian cycle from the Delamerian Orogen first and then from other orogens).

Delamerian cycle

While best developed in the Delamerian Orogen on the mainland and in west Tasmania, the Delamerian cycle is also represented in the early histories of the Lachlan, New England and North Queensland orogens. In west Tasmania, this cycle lasted *c.* 300 million years, from >780 Ma until 490 Ma, although it may have begun much earlier, depending on the tectonic significance of 1270 Ma and 1100 Ma ages below. On the mainland, it lasted *c.* 350 million years, from 830 Ma to 480 Ma.

Extension and passive margin phase

The oldest rocks in the Tasmanides occur in western Tasmania and on King Island between Tasmania and the mainland (Figs 3, 4a). They consist of:

- basalts in the west of King Island that underwent amphibolite-grade metamorphism at *c.* 1270 Ma (Holm *et al.* 2003);
- shallow-water undeformed sandstone in the NW corner of Tasmania with inferred *c.* 1100 Ma depositional ages (Turner *et al.* 1992).

Younger rocks that fit the rift history on the mainland better (see below) comprise:

- >780 Ma mafic volcanic rocks of the allochthonous Bowry Formation that underwent blueschist metamorphism (Holm *et al.* 2003). The Cooee Dolerite (with a minimum K–Ar age of 725 Ma, Crook 1979) is a correlative. Together they represent an early phase of rift volcanism (Holm *et al.* 2003);
- the Wickham deformation, dated by the emplacement of associated granites on King Island, at 760 Ma (Turner *et al.* 1998). This deformation also affected NW Tasmania, where similar granites are dated at 777 Ma (Turner *et al.* 1998);
- *c.* 700 Ma glacially-derived Sturtian and *c.* 635 Ma Marinoan conglomerates in

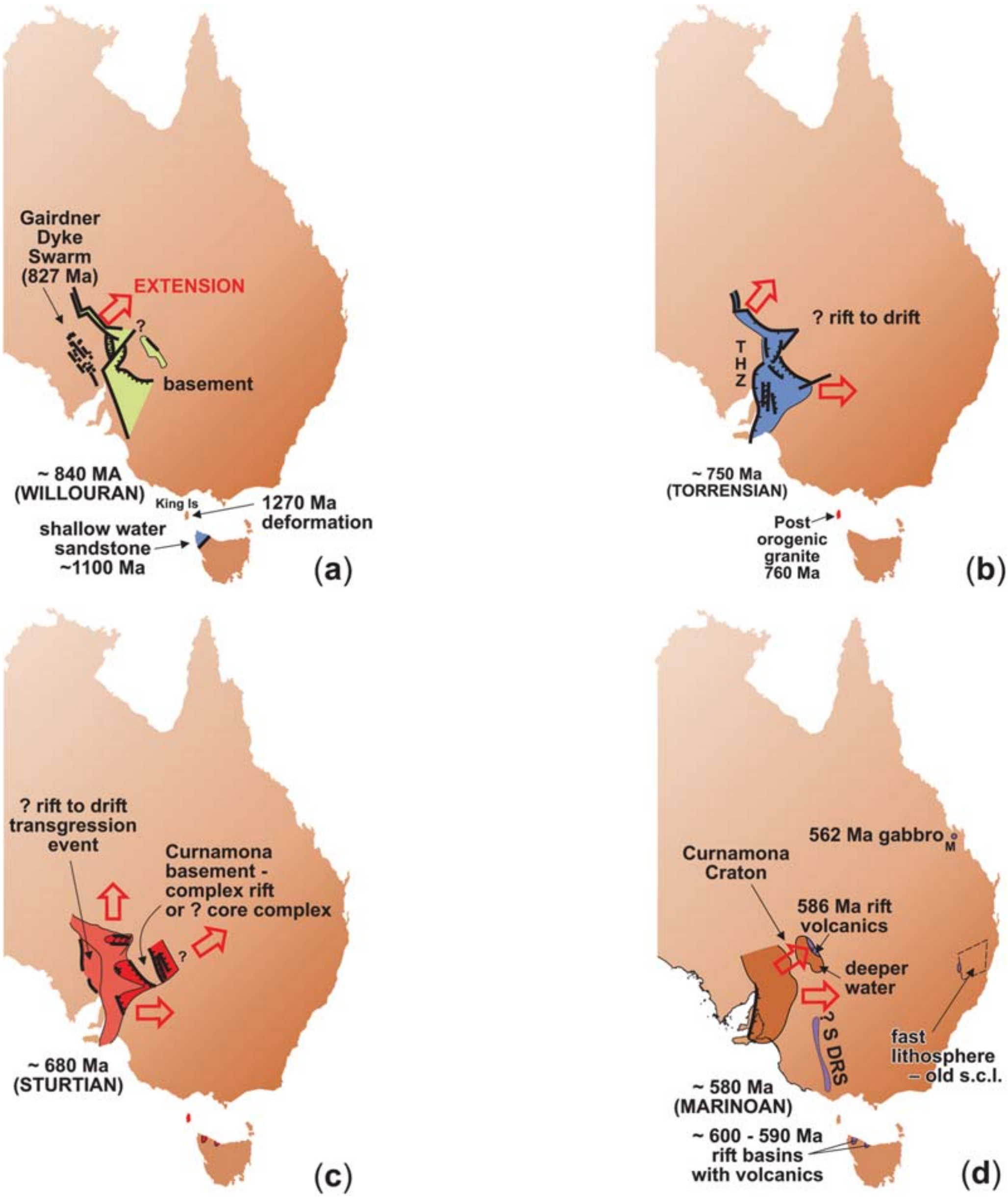


Fig. 3. Precambrian part of the Delamerian cycle. Rift cycle 1, Adelaide Rift Complex evolution, largely after Preiss (2000). Coloured polygons represent major depocentres and arrows represent inferred directions of maximum extension. (**a**) Willouran; (**b**) Torrensian (THZ, Torrens Hinge Zone); (**c**) Sturtian; (**d**) Marinoan (M, Marlborough; SDRS, seaward-dipping reflector sequences after Direen & Crawford (2003)), with inferred old subcontinental lithosphere (s.c.l.) under the New England Orogen.

younger basins in NW Tasmania (Figs 3c, d, 4) (Calver & Walters 2000);

- volcanic-rich rift basins, containing tholeiites passing up into second-stage melt lavas (picrites) (Crawford & Berry 1992) (Fig. 3d).

These igneous rocks formed during a major 600–580 Ma rifting event in western Tasmania and were interpreted as seaward-dipping reflector sequences by (Direen & Crawford 2003*a*) and (Crawford *et al.* 2003*a*).

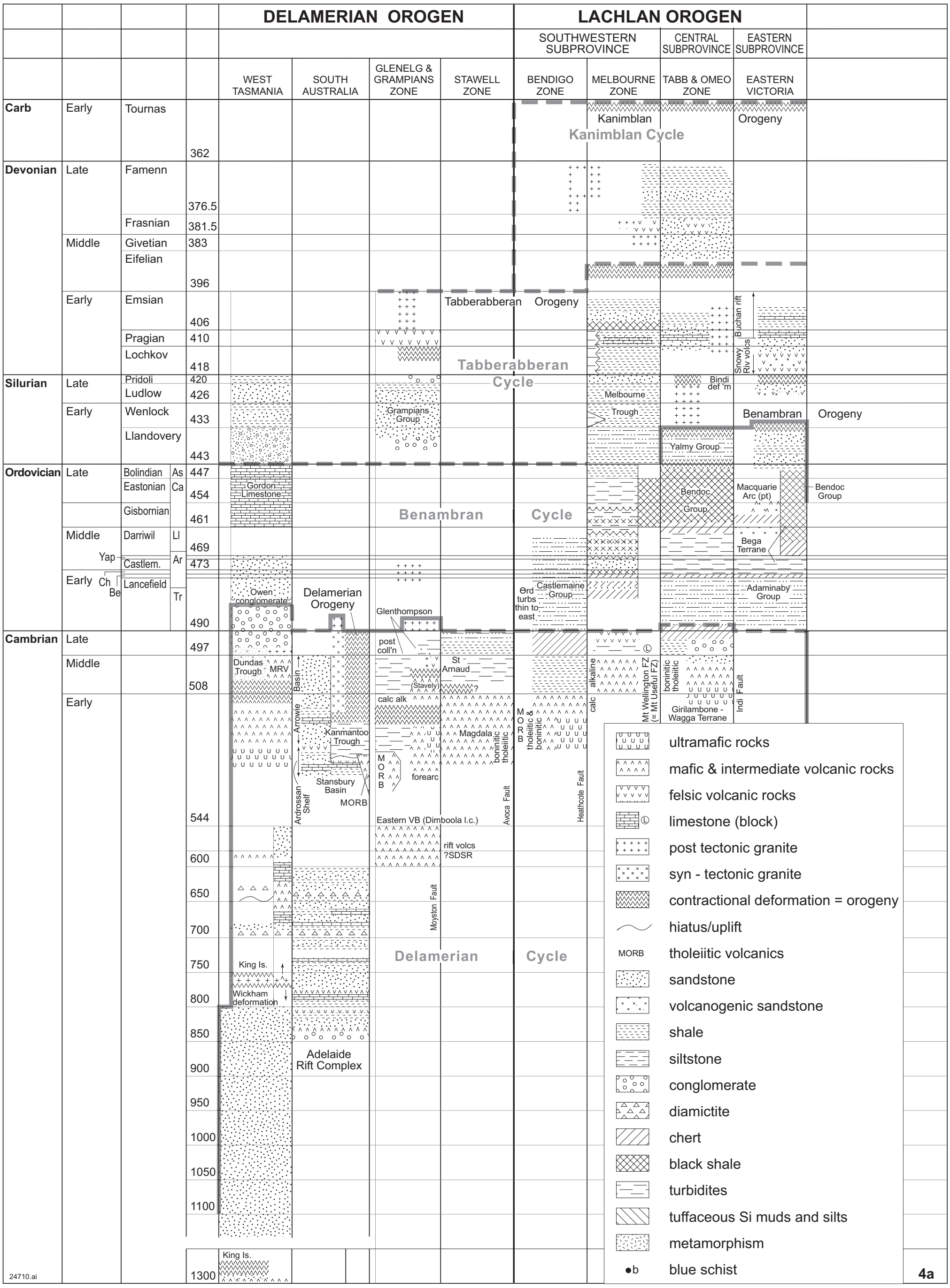


Fig. 4. Time–space plots. (**a**) Southern plot from Delamerian Orogen in South Australia and west Tasmania through to Lachlan Orogen in Victoria. Stratigraphic data mainly from Seymour & Calver (1995), Holm *et al.* (2003), Preiss (2000), VandenBerg *et al.* (2000) plus other sources cited in text. Time-scale – Veevers (2000*e*) is used as the source for most of the time-scale, with the exception of that part from the Ordovician to the Carboniferous which (rounded up or down) comes from Pogson & Percival (2003). For the Ordovician this latter scale is based on Cooper (1999). British stage names are also provided.

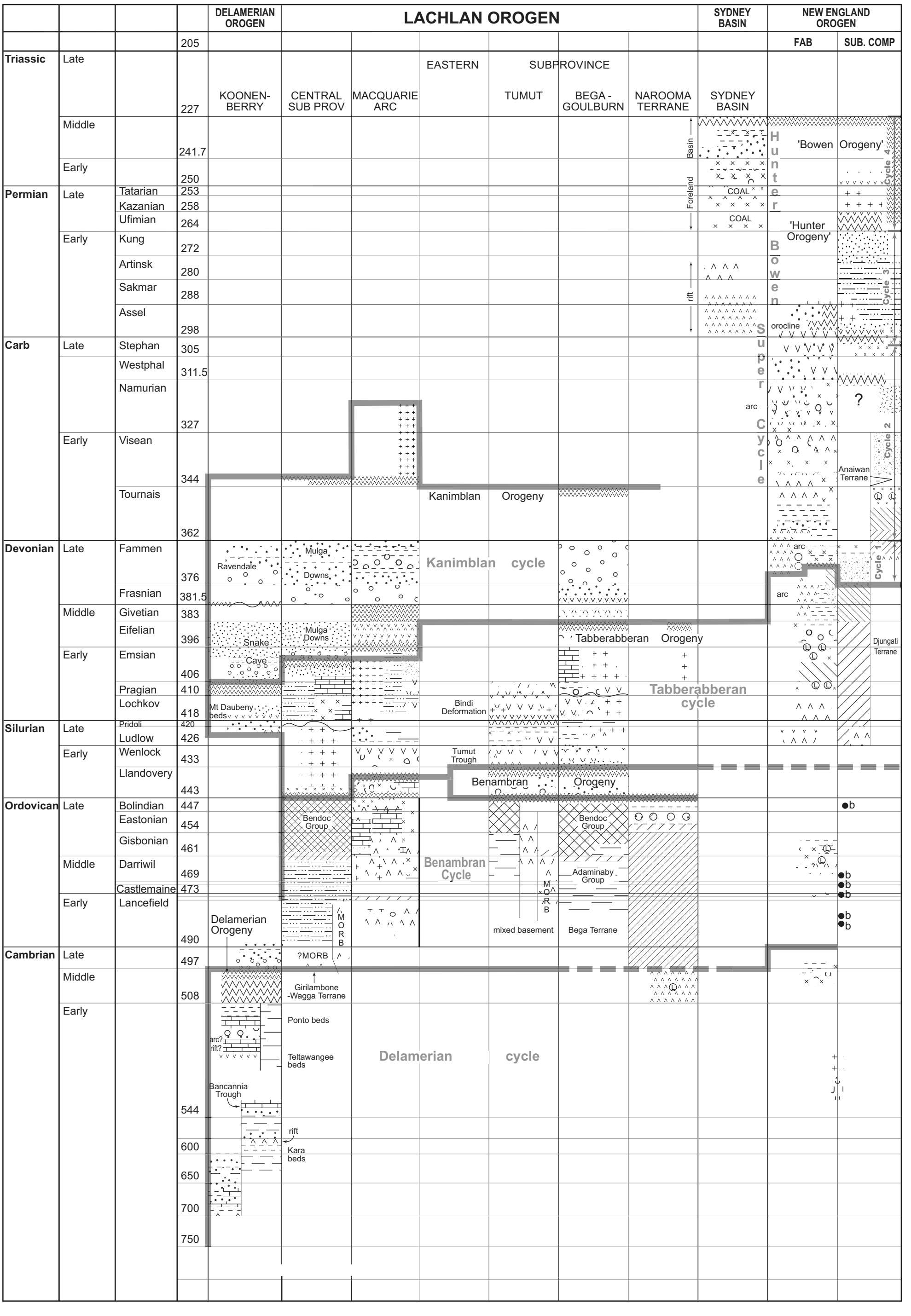


Fig. 4. (**b**) Central plot through New South Wales, from Delamerian Orogen (Koonenberry) through Lachlan Orogen into New England Orogen. Stratigraphic data mainly from Mills (2002, 2003), Colquhoun *et al.* (2004), Percival & Glen (2006), Meffre & Glen (unpublished), Thomas *et al.* (2002), Glen *et al.* (2004*a*), Roberts & Geeve (1999), Aitchison *et al.* (1992*a*), and other sources cited in text.

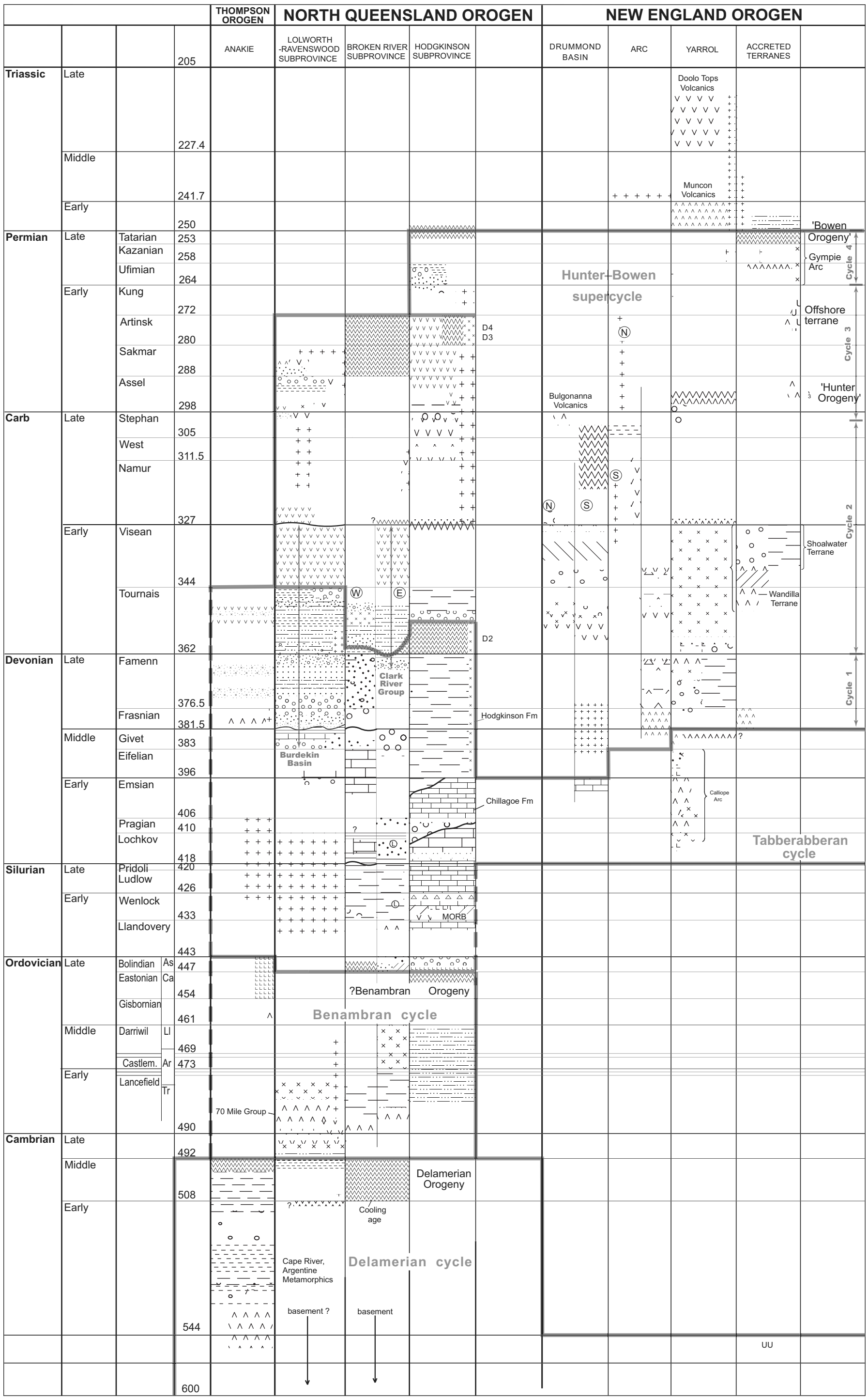


Fig. 4. (**c**) Northern plot through Queensland. Stratigraphic data mainly from Bain & Draper (1997), Bultitude *et al.* (1993), Withnall *et al.* (1996), Henderson *et al.* (1998), Leitch *et al.* (2003), the Yarrol Project Team (1997) and other sources cited in text.

On the mainland, there is no record of rifting before 830 Ma. The rift phase of the Delamerian cycle on the mainland is divided into two rift cycles. Rift cycle 1 is represented by growth of the Adelaide Rift Complex. Rift cycle 2 is represented by formation of the Kanmantoo Trough in eastern South Australia and equivalents in western Victoria and in the Koonenberry Belt of western New South Wales (Figs 1, 3, 5).

The Adelaide Rift Complex lies mainly west, but also east, of the Curnamona craton (Cooper & Tuckwell 1971). Its development records supercontinent break-up in the Neoproterozoic, beginning at 827 Ma and continuing to the base of the Cambrian. Five rifting events within this cycle have been identified by Preiss (2000) (Figs 3, 4a). In contrast to west Tasmania, rifting is essentially non-volcanic, although events 1 and 3 began with volcanism. U–Pb constraints on the ages of rifting are provided in the first event by the Wooltana Volcanics (*c.* 827 Ma, see Preiss 2000) coeval with the Gairdner Dyke Swarm to the west dated at *c.* 827 Ma by Wingate *et al.* (1998). Event 2 began with the Rook Tuff dated at *c.* 802 Ma (Fanning *et al.* 1986). Event 3 began with the rhyolites of the Boucaut Volcanics dated at *c.* 777 Ma (C. M. Fanning 1994, quoted by Preiss 2000). Event 4 began with Sturtian glacials, dated at *c.* 700 Ma and event 5 in the middle Marinoan at *c.* 650 Ma, below the Marinoan glacials dated at 635 Ma (Preiss 2000). Deeper-water equivalents (Kara beds) occur east of the Bancannia Trough in the Koonenberry Belt (Mills 1992, 2003).

A key feature of event 5 is the presence of alkaline rift volcanic rocks extruded over a large area from the Koonenberry Belt (where they are dated at *c.* 586 Ma, Crawford *et al.* 1997), southwards to the Truro alkaline volcanic rocks east of Adelaide, correlated by Crawford *et al.* (1997), and into the alkaline volcanic rocks of King Island and northwestern Tasmania. Also included in this event is the eastern 350 × 50 km belt of volcanic rocks in western New South Wales and Victoria concealed beneath Tertiary cover and revealed by aeromagnetic data (Figs 1, 5a). This belt, the Dimboola Igneous Complex of VandenBerg *et al.* (2000), contains ultramafic and mafic tholeiites, boninites, volcaniclastics and cumulate gabbros. They are interpreted as a set of *c.* 600 Ma seaward-dipping reflector sequences by Direen & Crawford (2003*a*). VandenBerg *et al.* (2000), however, suggested they represent a Cambrian intra-oceanic arc (see below).

Major tectonic changes occurred in the Middle Cambrian on the mainland (Figs 4a, 5) with the onset of rift cycle 2, which is absent in western Tasmania. On the mainland, a shallow-water Early Cambrian limestone shelf (part of the Stansbury Basin, Ardrossan Shelf in the west and Arrowie Basin in the north, e.g. Belperio *et al.* 1998; Preiss 2000) was developed above a regional hiatus on top of the western and northern parts of the Adelaide Rift Complex (Figs 4a, 5a). A tuff band at the top of the shelf sequence has a U–Pb date of *c.* 526 Ma (Cooper *et al.* 1992) (Fig. 4a). This shelf was followed by a major rifting event in the east (Waitpingan Subsidence, Thomson 1969; Haines & Flöttmann 1998) that led to the formation of a major new deep-water rift basin – the Kanmantoo Trough – that extended eastwards into western Victoria and has analogues in the Ponto beds of the Koonenberry Belt (Mills 2003).

The Kanmantoo Trough was filled by rapid deposition of *c.* 7–8 km thick (Jago *et al.* 2003), high-density turbidites (Haines *et al.* 2001) from 526 Ma till the onset of deformation around 514 Ma (Foden *et al.* 2002) (Fig. 4a). The Kanmantoo Trough formed as a transtensional basin in response to NE–SW extension, with W–E palaeocurrents reflecting deflection by bounding faults (Flöttmann *et al.* 1998; Haines & Flöttmann 1998).

The presence of basalts and gabbros, dated at *c.* 524 Ma and with within-plate and MORB chemistry (Rankin *et al.* 1991; VandenBerg *et al.* 2000), suggest that rifting had progressed to the formation of Cambrian oceanic crust.

The Kanmantoo Group was derived from Antarctica (Flöttmann *et al.* 1998) and shares the 600–500 Ma zircon population (the Pacific Gondwana zircons of Ireland *et al.* 1994) with Ordovician turbidites of the Lachlan Orogen, which are derived from the Ross Orogen of Antarctica (but see Williams *et al.* 2002). This source contrasts with derivation of the Adelaide Rift Complex (Turner *et al.* 1996; Ireland *et al.* 1998) and the partly coeval Early Ardrossan Shelf (Ireland *et al.* 1998), largely from the craton to the west (Veevers 2000*a*).

In the New England Orogen, the rift phase of the Delamerian cycle is represented by an ophiolite in the Marlborough Block of central Queensland that has a 562 ± 23 Ma Sm–Nd isochron age (Bruce *et al.* 2000) (Figs 3d, 4c). The ophiolite has depleted MORB-like trace element characteristics that suggest formation as oceanic crust at a Neoproterozoic ocean ridge (Bruce *et al.* 2000). These data are consistent with the existence of a proto-Pacific Ocean east of the Delamerian Orogen after supercontinent break-up.

In the western part of the Thomson Orogen, the Delamerian rift cycle is represented by continent-derived sandstones and mudstones with 580–480 Ma K–Ar deformation ages (Murray 1994; Scheibner & Veevers 2000). Correlation with rocks of rift cycles 1 or 2 to the southwest is uncertain. In the eastern part of the orogen, alkaline and tholeiitic volcanic rocks in the Anakie Inlier (Fig. 2) provide further evidence of rifting, while the presence of quartz sandstone suggests an intracontinental setting (Withnall 1995). These rocks may be as old as Neoproterozoic (Withnall *et al.* 1996), or late Neoproterozoic to Middle Cambrian (Fergusson *et al.* 2001).

The North Queensland Orogen includes inliers of metamorphic complexes, which appear to represent exhumed 1100–1200 Ma ('Grenvillian') continental crust (Blewett *et al.* 1998) as well as Neoproterozoic rift-related sediments and volcanic rocks indicative of an intraplate extensional origin (Draper *et al.* 1998; Hutton *et al.* 1998) (Figs 2, 4c). These inliers occur in both the Broken River and Lolworth–Anakie subprovinces, while the metamorphic rocks in the Barnard subprovince predate the intrusion of Ordovician granites, the oldest of

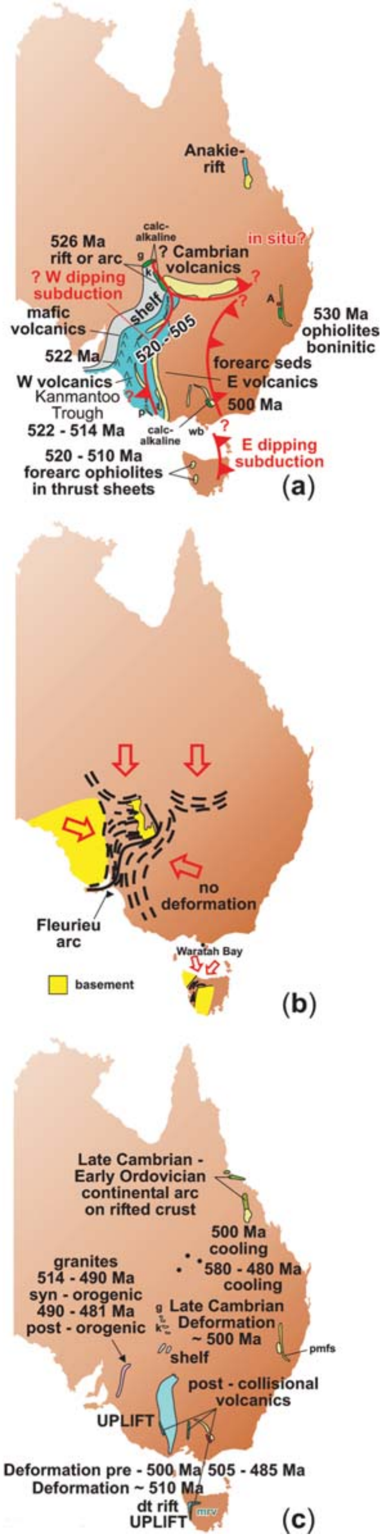


Fig. 5. Delamerian cycle, rift cycle 2, convergent, collisional and post-collisional phases. (**a**) Blue unit is *c.* 522–514 Ma Kanmantoo Trough with mafic volcanic rocks (^). Pale green units are deformed volcanic packets. Eastern volcanics = Dimboola Igneous Complex – either a Cambrian arc or a 600–580 Ma seaward-dipping reflector sequence (see Fig. 3d). Subduction zones shown in red with red barbs. East-dipping subduction zone based on model of Crawford & Berry (1992). Possible north- and west-dipping subduction zones postulated in this paper based on the presence of inferred Cambrian volcanic rocks. Eclogite at Attunga (A) in northern NSW and 530 Ma ophiolites along the Peel–Manning Fault System (pmfs) are also part of the convergent phase. (**b**) Collisional phase, showing regional trends in black wrapping around Precambrian continental buttresses shown in yellow. Arrows show directions of maximum shortening. Fleurieu structural arc highlighted. (**c**) Syn- to post-collisional phase showing syn- and post-orogenic granites and post-collisional volcanic rocks and sedimentary rocks. Dark blue area represents post-collisional turbidites; light blue areas represent post-collisional shallow-water deposits. Abbreviations: A, Attunga; dt, Dundas Trough; g, Gidgealpa; k, Koonenberry; mrv, Mt Read Volcanics; p, Padthaway; pmfs, Peel–Manning Fault System; wb, Waratah Bay.

which is 486 Ma (Bultitude & Garrad 1997) (Fig. 2). These inliers may be correlatives of the Anakie Inlier (Withnall 1995), since they predate Delamerian deformation and cooling (Rb–Sr isochron at *c.* 500 Ma from the Cape River Metamorphics in the Broken River Subprovince, Draper *et al.* 1998) (Fig. 5a). In the Cape River Metamorphics, amphibolite-grade sandstones with *c.* 1145 Ma dominant detrital zircons are intruded by Late Cambrian–Early Ordovician granite and overlain by Late Cambrian–Early Ordovician volcanic rocks (see below) (Hutton *et al.* 1998). In the Lolworth–Ravensworth Block, the Charters Towers Metamorphics contain *c.* 507 Ma mafic volcanic rocks (Bain & Draper 1997). The presence of a Grenvillian basement is supported further by Neoproterozoic detrital zircons in Palaeozoic strata, old zircons in early Palaeozoic granites and, significantly, negative epsilon Nd values (Bain & Draper 1997).

Convergent phase

The convergent phase of the Delamerian cycle is represented by development of crust formed in the forearc of an intra-oceanic island arc.

Convergence between the extended west Tasmania craton and the proto-Pacific plate is reflected by the development of Cambrian mafic–ultramafic complexes. These constitute relics of forearc igneous crust that developed on the proto-Pacific plate above an east-dipping subduction zone and west of an intra-oceanic arc (Crawford & Berry 1992). Reed *et al.* (2002) suggested that the east-dipping seismic reflectors of Barton (1999) in the west Tamar region reflect the location of the Cambrian Delamerian subduction zone (Fig. 5a). A zircon date *c.* 510 Ma from the Heazlewood Complex (Turner *et al.* 1998) reflects ongoing subduction and dates convergence from *c.* 520 to 510 Ma (A. Crawford pers. comm. 2004) (Fig. 4).

Candidate arc-rocks on the mainland are concealed beneath Mesozoic and younger cover, but are revealed by geophysical data. Two NSW–NNE-trending belts, an eastern and western, each *c.* 350 × 50 km, occur in western New South Wales, Victoria and eastern South Australia. The third, a curved east–west belt, lies along the southern margin of the Thomson Orogen (Figs 1, 5a).

The eastern belt is the Dimboola Igneous Complex of VandenBerg *et al.* (2000), interpreted as 580–600 Ma rift volcanic rocks by Direen & Crawford (2003*a*) but as a Cambrian intra-oceanic arc reflecting convergence along the Gondwana proto-Pacific plate boundary by VandenBerg *et al.* (2000). Elements of rift, forearc crust and post-collisional volcanic rocks may be present since the small Stavely Volcanic Complex just to the east and south contains slices of serpentinized boninitic ultramafics, interpreted as remnants of the forearc of an intra-oceanic island arc, overlain by post-collisional felsic volcanic rocks (Crawford *et al.* 1996) (Fig. 4a). The tholeiitic Magdala Volcanics in the Stawell Zone lie immediately east of the Dimboola Igneous Complex and the Moyston Fault. Although equivalent in age, they lack the strong geophysical response on a regional scale. Crawford *et al.* (2003*b*) quoted a mixing age of 518 Ma.

The western volcanic belt in the Delamerian Orogen (Figs 1, 5a) wraps around the Precambrian Curnamona craton and contains a mixture of back-arc, arc-like and ?post-collisional volcanic rocks with zircon ages of 521–480 Ma. The Padthaway area (Fig. 1) contains MORB volcanic rocks (Rankin *et al.* 1991) and felsic volcanic rocks intruded by syn- to post-deformation mafic and granitic intrusive rocks. One felsic volcanic rock has been U–Pb dated at *c.* 493 Ma, and one granite at *c.* 480 Ma (Fanning 1996). Gabbro (just east of the P, Fig. 1) has been U–Pb dated at 525 Ma (Maher *et al.* 1977). Further north, in the ENE part of the belt, Neoproterozoic and Cambrian mafic volcanic and sedimentary rocks are intruded by Cambrian–Ordovician, late- to post-deformation granitoids and mafic to intermediate igneous rocks such as microdiorites (U–Pb dated at *c.* 482 Ma, Fabris 2003). Just ESE of Broken Hill, K-rich volcanic to subvolcanic rocks are intruded by granite, diorite and monzodiorite U–Pb dated at 521 Ma, 519 Ma and 505 Ma (cited by Mills 2001).

The Koonenberry area near the northern part of this western belt (Fig. 5a) contains calc-alkaline andesitic volcanic rocks overlain by late Early Cambrian felsic volcanic rocks U–Pb dated at 525 Ma (J. Claoué-Long 1992, cited in Zhou & Whitford 1994) below Middle Cambrian limestone, shale and volcanolithic conglomerate. These shallow-water rocks pass eastwards into fault-bounded turbidites of the Teltawongee beds and overlying Ponto beds that are equivalent to the Kanmantoo Group (Mills 1992). Concealed Early Cambrian volcanic rocks at Gidgealpa at the northern tip of the western belt consist of a trachytic lower part and an upper part of rhyolitic to dacitic tuff beds with some andesites (Gatehouse 1986; Gravestock & Gatehouse 1995) (Fig. 5a).

The calc-alkaline volcanic rocks at Koonenberry were interpreted as an arc by Scheibner (1987) and Scheibner & Basden (1998), although Zhou & Whitford (1994) and Crawford *et al.* (1997) suggested that they represented rift volcanism from their geochemistry, their isotope contents and, in part, also from the small amount of volcanic material. A rift setting was also ascribed to the Gidgealpa Volcanics by Gravestock & Gatehouse (1995). However, Sharp & Buckley (2003) suggested that the Koonenberry rocks are subduction-related and included with them 520 Ma volcanic rocks intersected in drill holes beneath the Bancannia Trough just to the west.

In the Lachlan Orogen, elements of Delamerian convergence are represented by narrow zones of Cambrian mafic and ultramafic rocks in the Southwestern subprovince (Figs 5a, 6). These Cambrian igneous rocks are exposed mainly as fault slices in the hangingwalls of major thrust faults (Fig. 6): Heathcote Fault zone (the western edge of the Melbourne Trough), Mt Wellington Fault zone (the eastern edge of the Melbourne Trough) and the Governor Fault zone (the western edge of the Tabberabbera Zone). They also occur as very small bodies within the Bendigo and Melbourne structural zones (Ceres, Phillip Island, Waratah Bay – only the last is shown in Fig. 5).

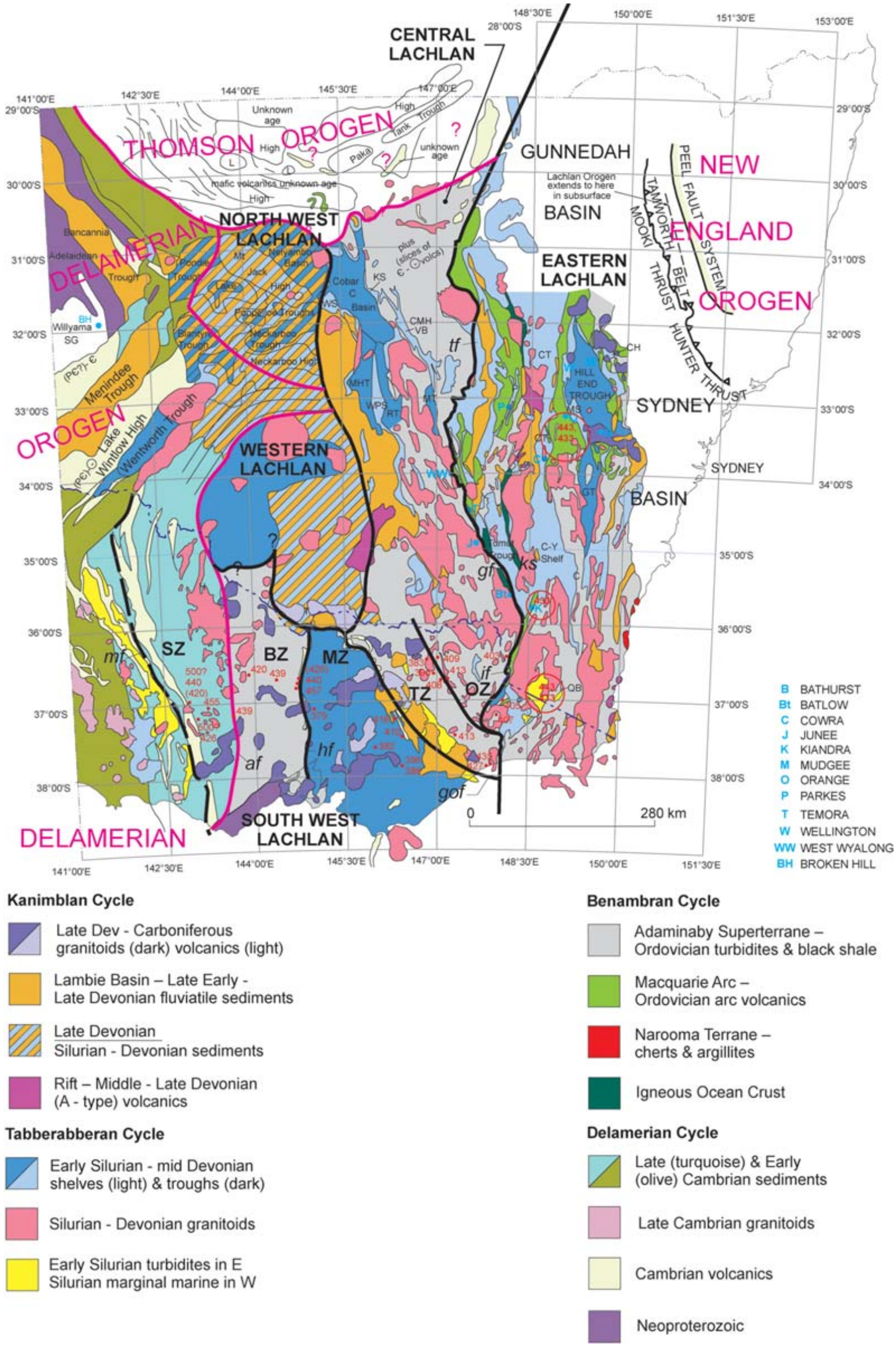
Most of these igneous rocks are boninitic andesitic lavas and ultramafic equivalents of forearc affinity, as well as tholeiitic basalt, the latter with back-arc basin geochemical signatures that may reflect rifting of the forearc (Crawford *et al.* 1984; Crawford *et al.* 2003*b*). A gabbro (at Dookie) at the northern end of the Governor Fault zone was dated at 502 Ma by Spaggiari *et al.* (2003*a*). In contrast, calc-alkaline andesites occur along, and in windows below, the Mt Wellington Fault zone (renamed the Mt Useful Fault zone by VandenBerg *et al.* 2000). These andesites pass up into volcaniclastic rocks and are overlain by sandstones and cherts. One body (at Licola in the south) has been dated by U–Pb at 500 Ma (Spaggiari *et al.* 2003*a*).

The third volcanic belt, largely concealed, lies in the southern margin of the Thomson Orogen and may also contain Cambrian volcanic rocks of the convergent phase of the Delamerian cycle. The presence of a significant gravity high and several magnetic ridges suggests that this margin of the orogen contains curvilinear east–west-trending major igneous bodies (Fig. 5a). Drill hole intersections (in the southern high near longitude 145°30′ S (Fig. 6) indicate the presence of gabbro, andesitic volcanic and volcaniclastic rocks, amphibole peridotites and sediments (Savage 2000), but there are few data as to the nature or ages of these hidden bodies. Scheibner & Basden (1998) inferred a Devonian rift, but the presence of Late Ordovician graptolites (summarized in Glen *et al.* 1996) and Cambrian to Early Ordovician zircons in the Easter Monday beds north of Koonenberry (Stevens 1991; Stevens & Fanning unpublished) indicates that they are older. If they predate Delamerian north–south shortening in the Koonenberry Belt (Mills 2003), they could represent a mixture of Cambrian rift and convergent margin igneous rocks and ultramafics along the southern margin of the Thomson Orogen.

In the southern New England Orogen, the convergent part of the Delamerian cycle is represented by eclogite blocks, exhumed in a major serpentinite-lubricated fault at Attunga on the Peel–Manning Fault System (Fig. 5a). Originally dated at 571 Ma by Watanabe *et al.* (1998), this date has now been revised to 536 Ma by Fanning *et al.* (2002) and is similar to the 530 Ma zircon ages obtained from plagiogranites and metadiorites in schistose serpentinite along the same fault system (Figs 4b, 5a) (Aitchison *et al.* 1992*b*; Sano *et al.* 2004 respectively). The eclogite indicates Cambrian subduction. The plagiogranite ages suggest that the enclosing ophiolitic low-Ti tholeiitic basalts and boninitic ultramafic rocks are relics of a Cambrian suprasubduction zone forearc and, thus, of a Cambrian convergent plate boundary (Aitchison *et al.* 1994). Chemically, the basalts resemble Cambrian basalts of western Tasmania (Aitchison & Ireland 1995). Convergence is also indicated by Middle to early Late Cambrian volcaniclastic rocks of the Murrawong Creek Formation that occur immediately west of the Peel–Manning Fault System (Figs 4b, 5a), dated from fossils in limestone clasts (Cawood 1976; Cawood & Leitch 1985). Although Aitchison & Flood (1990) suggested that Cambrian fossils came from allochthonous blocks in Devonian matrix, the conformably overlying Pipeclay Argillite contains Middle to early Late Cambrian conodonts (Stewart 1995). Cawood & Leitch (1985) suggested that volcanic clasts in the conglomerate were derived from a low-K intra-oceanic island arc.

Collisional phase

Western Tasmania records the collision of the forearc of an intra-oceanic arc with extended East Gondwana crust around 510–505 Ma. This resulted in the southwest transport of allochthonous thrust sheets of forearc mafic and



CENTRAL LACHLAN
THOMSON OROGEN
GUNNEDAH BASIN
NEW ENGLAND OROGEN
PEEL FAULT SYSTEM
TAMWORTH – BELT
MOOKI THRUST
HUNTER THRUST
Lachlan Orogen extends to here in subsurface
NORTH WEST LACHLAN
EASTERN LACHLAN
DELAMERIAN OROGEN
WESTERN LACHLAN
HILL END TROUGH
SYDNEY
SYDNEY BASIN
Menindee Trough
Lake Wintlow High
Wentworth Trough
Cobar Basin
SZ
BZ
MZ
TZ
OZ
SOUTH WEST LACHLAN
DELAMERIAN
0
280 km
B BATHURST
Bt BATLOW
C COWRA
J JUNEE
K KIANDRA
M MUDGEE
O ORANGE
P PARKES
T TEMORA
W WELLINGTON
WW WEST WYALONG
BH BROKEN HILL
Kanimblan Cycle
Late Dev - Carboniferous granitoids (dark) volcanics (light)
Lambie Basin – Late Early - Late Devonian fluviatile sediments
Late Devonian
Silurian - Devonian sediments
Rift – Middle - Late Devonian (A - type) volcanics
Tabberabberan Cycle
Early Silurian - mid Devonian shelves (light) & troughs (dark)
Silurian - Devonian granitoids
Early Silurian turbidites in E Silurian marginal marine in W
Benambran Cycle
Adaminaby Superterrane – Ordovician turbidites & black shale
Macquarie Arc – Ordovician arc volcanics
Narooma Terrane – cherts & argillites
Igneous Ocean Crust
Delamerian Cycle
Late (turquoise) & Early (olive) Cambrian sediments
Late Cambrian granitoids
Cambrian volcanics
Neoproterozoic

ultramafic rocks over the thinned passive margin sequence and its cover of late Neoproterozoic rift basins (Berry & Crawford 1988; Crawford & Berry 1992; Berry 1994; 1995) (Figs 5a, 8). Ultramafic detritus in the middle Middle Cambrian basal units of the Dundas Trough indicates that obduction (Fig. 5b) had occurred between latest Early Cambrian and middle Middle Cambrian. Obduction was accompanied by the formation of metamorphic complexes beneath the ophiolitic sheets (Meffre *et al.* 2000), with one eclogite U–Pb dated at 502 Ma (Turner *et al.* 1998).

Subsequent deformation between 505 Ma and 495 Ma is subdivided into an early phase of N–S compression in the late Middle–early Late Cambrian and a later phase of E–W compression in the Late Cambrian (Berry 1994; Turner *et al.* 1998). The high-strain Arthur Lineament running NE across western Tasmania is interpreted by Holm & Berry (2002) as a series of thrust sheets emplaced N–S in the first event and refolded and steepened by subsequent E–W folding and faulting. Woodward *et al.* (1993) showed that allochthonous Precambrian massifs were also emplaced in the Late Cambrian.

On the mainland, the Delamerian Orogeny produced an orogenic belt between 300 km and 600 km wide (Fig. 5b). The western external part of this orogen is marked by a 100–300 km wide fold–thrust belt in the external part of the orogen in the west, and an extensive, internal high-T low-P zone extending eastwards under the Cenozoic Murray Basin into western Victoria. This zone is marked by multiple deformation and metamorphism and the emplacement of syn- and post-kinematic granites. Low-grade (dominantly volcanic) rocks also occur along the eastern margin of the orogen in western Victoria, suggesting that the Delamerian Orogen may have lower-grade margins flanking a higher-grade core.

The western external fold–thrust belt has variable geometry. The central and northern parts were deformed by south-vergent thrusting that is thin skinned, except where evaporites of

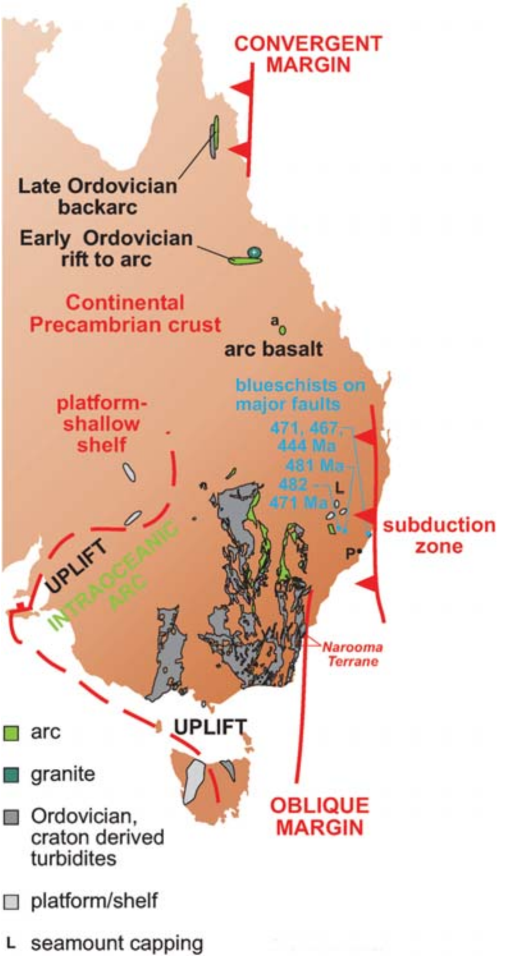
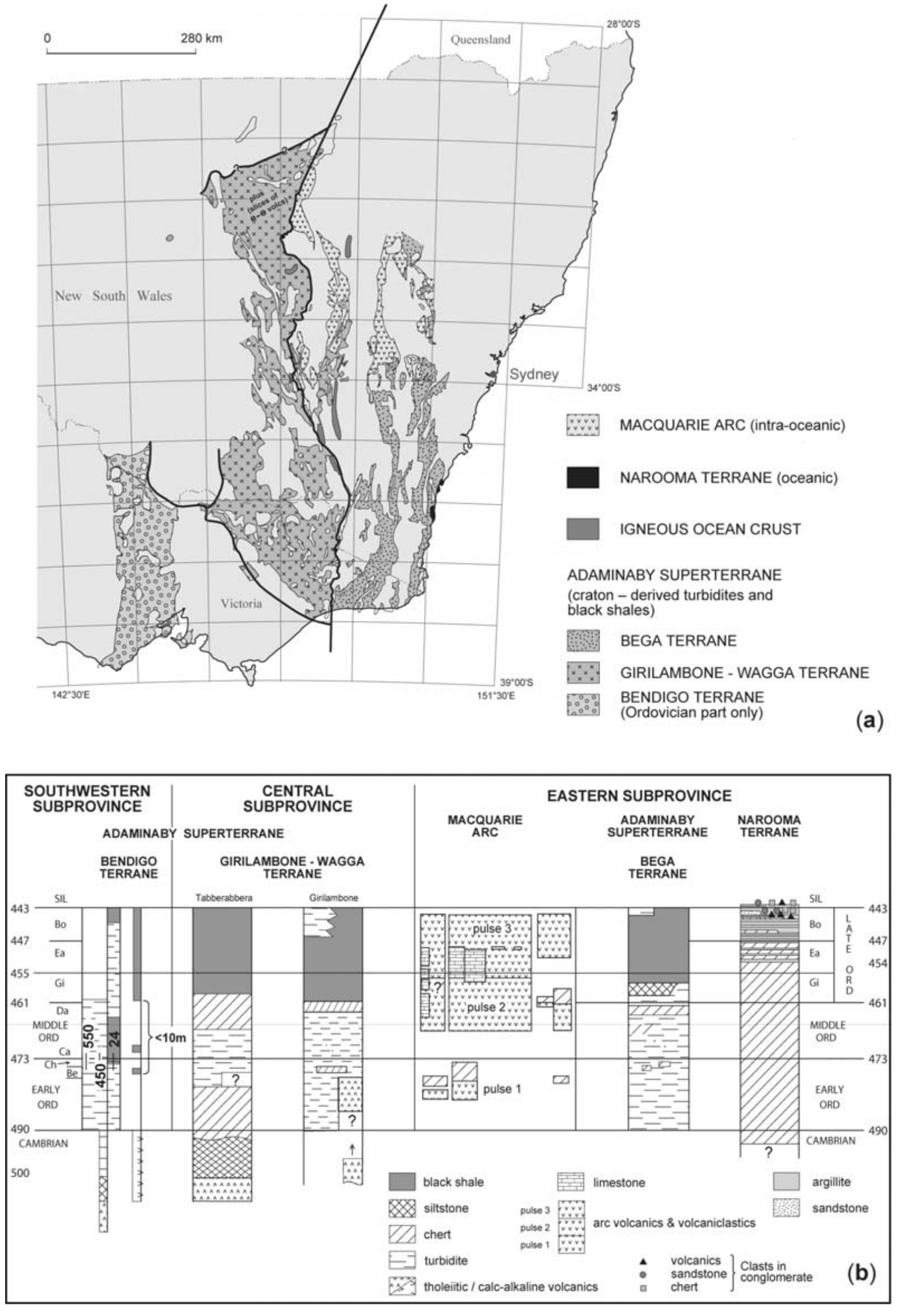


Fig. 7. Benambran cycle showing craton-derived turbidite, arc and Narooma terranes in Lachlan Orogen, blueschist ages in New England Orogen (from text) and inferred tectonic elements in northern New England Orogen and in North Queensland Orogen. Platform deposits and uplift in Delamerian orogen to west also shown. P, Port Macquarie; a, location of Anakie Inlier.

Fig. 6. Map of Lachlan Orogen showing subdivision into subprovinces, major faults and tectonostratigraphic units. Red numbers are Ar–Ar plateaux ages from Foster *et al.* (1999). Blue numbers are deformation ages from the stratigraphic record (Glen *et al.* 2004*b*). Highs and lows (L) in the southern Thomson Orogen refer to gravity features. Abbreviations: SZ, Stawell Zone; BZ, Bendigo Zone; MZ, Melbourne Zone; TZ, Tabberabbera Zone; OZ, Omeo Zone: mf, Moyston Fault; af, Avoca fault; hf, Heathcote Fault; gof, Governor Fault Zone; if, Long Plain & Indi faults; gf, Gilmore Fault Zone; tf, Tullamore Fault; ks, Kiandra–Narromine Structure; CT, Cowra Trough; CH, Capertee High; WS, Winduck Shelf; KS, Kopyje Shelf; CMH VB, Canbelego–Mineral Hill Volcanic Belt; MHT, Mt Hope Trough; WPS, Walters Range Shelf; RT, Rast Trough; MT, Melrose Trough; QB, Quidong Basin; C-Y Shelf, Canberra-Yass Shelf. Geology in Victoria based on VandenBerg *et al.* (2000).


0
280 km
Queensland
28°00'S
New South Wales
Sydney
34°00'S
Victoria
39°00'S
142°30'E
151°30'E
MACQUARIE ARC (intra-oceanic)
NAROOMA TERRANE (oceanic)
IGNEOUS OCEAN CRUST
ADAMINABY SUPERTERRANE
(craton – derived turbidites and black shales)
BEGA TERRANE
GIRILAMBONE - WAGGA TERRANE
BENDIGO TERRANE
(Ordovician part only)
(a)
SOUTHWESTERN SUBPROVINCE
CENTRAL SUBPROVINCE
EASTERN SUBPROVINCE
ADAMINABY SUPERTERRANE
BENDIGO TERRANE
GIRILAMBONE - WAGGA TERRANE
Tabberabbera
Girilambone
MACQUARIE ARC
ADAMINABY SUPERTERRANE
BEGA TERRANE
NAROOMA TERRANE
SIL
443
Bo
447
Ea
455
Gi
461
Da
MIDDLE ORD
Ca
473
Ch
Be
EARLY ORD
490
CAMBRIAN
500
550
450
24
<10m
pulse 3
pulse 2
pulse 1
LATE ORD
454
black shale
siltstone
chert
turbidite
tholeiitic / calc-alkaline volcanics
limestone
arc volcanics & volcaniclastics
volcanics
sandstone
chert
Clasts in conglomerate
argillite
sandstone
(b)

the Callana Group are absent (Preiss 2000) (Fig. 5b). The southern part of the fold-and-thrust belt verges westwards towards the Gawler craton and outlines the Fleurieu structural arc (Fig. 5b), which reflects variable shortening against the Gawler craton buttress (Flöttmann *et al.* 1994; Flöttmann & James 1997). The thrust belt involved only very limited translation, since thrusts do not extend west of the Torrens Hinge Zone – the western edge of the Adelaide Rift Complex from Sturtian times onward. This is consistent with formation of a very restricted foreland basin that lies immediately west of that fault zone (Flöttmann *et al.* 1997). Haines & Flöttmann (1998) also suggested that some Cambrian sediments east of the Torrens Hinge Zone were deposited in a foreland basin that began at *c.* 523 Ma, but this is much older than the onset of deformation in the internal zone at 514 Ma (Foden *et al.* 2002). It is, however, consistent with a 531 Ma Rb–Sr age of cleavage formation in one part of the thrust belt (Turner *et al.* 1994).

In the internal part of the orogen, the Kanmantoo Trough underwent severe inversion during this Delamerian Orogeny, undergoing multiple deformation, metamorphism and granite emplacement (e.g. Jenkins & Turner 1992) (Fig. 4a). Old growth faults were reverse-reactivated and there was major thrusting (Flöttmann & James 1997; Flöttmann *et al.* 1998). Syn-kinematic granite ages indicate that deformation commenced at 514 Ma and lasted till 490 Ma (Foden *et al.* 2002). Subsequent intrusive rocks are largely post-kinematic, high-level A-type granites and gabbros (497–481 Ma, Foden *et al.* 2002) that were emplaced during 15 km of exhumation (S. P. Turner *et al.* 1992).

In the outboard part of the orogen, in the Koonenberry area, deformation was somewhat younger, dated as Middle to Late Cambrian (Mills 2003). Neoproterozoic rocks west of the Bancannia Trough were deformed into west-verging structures (Cooper *et al.* 1975). East of the trough, turbidites were juxtaposed against shallow-water limestones and volcanic rocks. Increasing deformation in Cambrian rocks to the northeast (Mills 2002) implies a component of N–S shortening (Fig. 5b). Direen (1997) and Direen & Crawford (2003*b*) used geophysical data to infer a stack of west-vergent thrusts, but more recent interpretation of deep seismic data suggests major west-dipping structures and a major crustal antiform, with an amplitude of *c.* 15 km and cored by Neoproterozoic strata (Mills & David 2003).

In the outboard part of the orogen, in western Victoria, volcanic rocks thrust westwards over the Kanmantoo Group equivalents and older packages are interpreted to be either forearc crust (Crawford *et al.* 1996) or an arc (Scheibner 1989; VandenBerg *et al.* 2000) (Fig. 6). Thrusting predated the 500 Ma post-collisional volcanic rocks in the Mount Stavely Volcanic Complex (Crawford *et al.* 1996) (Fig. 4a). Most of the deformation (D1–D5) and high-T low-P heating were taken up by the sediments in the collision zone west of the volcanic rocks, where a 40 km wide high-grade zone containing syn- and post-tectonic granites and migmatites passes across faults into lower-grade sedimentary and volcanic rocks to the east and west (e.g. Gibson & Nihill 1992; Gray *et al.* 2002; Kemp *et al.* 2002). Peak deformation and metamorphism occurred around 516 Ma (Turner *et al.* 1993), with cooling ages of 500–480 Ma from metamorphic rocks and granites (Richards & Singleton 1981; Turner *et al.* 1993).

Amphibolite-grade rocks occur further east in the Stawell Zone, in the hanging wall of the west-vergent Moyston Thrust (Fig. 6), which has undergone vertical displacement of 15–20 km (Phillips *et al.* 2002). Miller *et al.* (2003) reported that mineral growth occurred at *c.* 500 Ma in the high-grade rocks, thereby indicating that the Moyston Fault was a Delamerian structure (compare Taylor & Cayley 2000 and Gray & Foster 2000). East of the Moyston Thrust, metamorphic grade drops back to low-grade greenschist facies, with most of the Stawell Zone cut by east-vergent thrusts (VandenBerg *et al.* 2000). The Stawell Zone lacks clear-cut evidence of a Delamerian unconformity.

In the Lachlan Orogen, Delamerian (or Tyennan) deformation has been inferred in one locality only – at Waratah Bay (Fig. 5b) – where an unconformity separates basalt from chert (Cayley *et al.* 2002). Spaggiari *et al.* (2003*c*) suggested, however, that this discordance could represent local features on a topographic high rather than part of a major orogenic event. Conformable Cambrian–Ordovician relationships are, thus, the inferred norm, based on relations in:

Fig. 8. Ordovician terranes in the Lachlan Orogen: (**a**) distribution; (**b**) time–space plot. In the Bendigo terrane, the thickness of the Bendigonian 4 to Castlemainian 1 part of the package thins eastwards from 550 m in the west through 450 m down to 24 m in the southwestern corner of the Melbourne Zone (central column). In the eastern part of the Melbourne Zone (right-hand column) the Bendigonian to Darriwilian sequence is <10 m thick. Thicknesses from VandenBerg *et al.* (2000). Other sources cited in text.

1. the hanging wall of the Heathcote Fault zone, at Lancefield (Cas & VandenBerg 1988; VandenBerg *et al.* 1992);
2. the hanging wall of the Governor Fault in the upper Howqua River (Crawford 1988; Fergusson 1998), where Cambrian volcanic and volcaniclastic rocks pass up into clastic rocks and then Late Cambrian chert, below thick Early Ordovician sandstone;
3. in the Narooma Terrane, where Late Cambrian and Early Ordovician cherts appear to be conformable (Glen *et al.* 2004*a*).

These observations imply that the Delamerian/Tyennan deformation did not extend into the Central or Eastern subprovinces of the Lachlan Orogen (but see Part 3). Nor is there any record in the New England Orogen (Cawood & Leitch 1985).

The Delamerian Orogeny affected the eastern part of the Thomson Orogen, the Anakie Inlier. Rocks here underwent a *c.* 500 Ma K–Ar cooling that postdates formation of a shallow S2 cleavage (Withnall *et al.* 1996) (Fig. 5c). Fergusson *et al.* (2001) suggested that S1 occurred at *c.* 583 Ma and S2 at *c.* 540 Ma.

In the North Queensland Orogen, the Late Cambrian–Early Ordovician Seventy Mile Group (Henderson 1986) described below from the Benambran cycle along the southern margin of the orogen shows no effects of any Delamerian Orogeny. However, the presence of *c.* 500 Ma cooling ages just to the north in basement inliers (the Cape River Metamorphics and Barnard Metamorphics; e.g. Draper *et al.* (1998)) implies either that deformation predated the Late Cambrian or that some units are allochthonous.

Post-collisional phase

Collision in west Tasmania was followed by extension and dismemberment of the thickened crust, leading to formation of the Dundas Trough and the Mount Read Volcanics to the east (Figs 4a, 5c) (Crawford & Berry 1992). These volcanic rocks were erupted between 505 Ma and 495 Ma (Perkins & Walshe 1993) and intruded by late dacites and granites. Continued rifting led to exhumation of the eastern basement and opening of rift basins, filled first by the Owen Conglomerate, with local internal unconformities that date deposition from *c.* 490 to 470 Ma (Seymour & Calver 1995), passing up into sandstone and then a sag phase platform sequence of Ordovician limestone (Noll & Hall 2003; 2004) (Fig. 4a). Deformation phases of the Delamerian/Tyennan Orogeny continued into the earliest Ordovician (Seymour & Calver 1995; Holm *et al.* 2003).

In the outboard part of the orogen, in the Koonenberry area, post-collisional units consist of basal conglomerate passing up into shallow-water to fluviatile Late Cambrian–Early Ordovician sediments (Mootwingee and Kayrunnera groups) (Mills 2002; 2003) (Fig. 5c). To the north at Gidgealpa, equivalent 'molassic' strata form the lower part of the Dullingari Group, which extends into the Late Ordovician (Gravestock & Gatehouse 1995).

Further south in the outboard part of the orogen in Victoria, accretion of forearc (and arc) volcanics was followed by rapid extension, reflected by:

1. the emplacement of calc-alkaline post-collisional volcanic rocks into the accreted arc (upper part of the Mount Stavely Volcanic Complex) (Crawford *et al.* 1996);
2. formation of accommodation space for accumulation of turbidites of the Glenthompson Sandstone and in the Stawell Zone, the 2–2.5 km thick St Arnaud Group of inferred Middle–Late Cambrian age above the Magdala Volcanics (VandenBerg *et al.* 2000; I. Williams cited in VandenBerg *et al.* 2000; Squire *et al.* 2003) (Figs 4a, 5c, 6); and
3. formation of extensional faults (Cayley & Taylor 1996; VandenBerg *et al.* 2000).

The post-collisional turbidites were themselves deformed in the Late Cambrian, in a second phase of the Delamerian Orogeny, which is constrained by post-tectonic granites, the oldest of which has a U–Pb age of 489 Ma (Crawford *et al.* 1996; VandenBerg *et al.* 2000). The vast amounts of detritus shed eastwards from the eastern part of the orogen suggests that deformation here was also east directed, producing a divergent orogen.

Lachlan supercycle

The Lachlan supercycle characterizes the Ordovician to Carboniferous history of the Lachlan Orogen and lasts for *c.* 170 million years, from *c.* 490 Ma to 320 Ma. This supercycle is divided into three cycles, separated by major contractional deformations that appear to be orogen-wide in space and time, although commonly they may be multiphase, diachronous in detail and variable in intensity. These three cycles are the Benambran (*c.* 50 million years, 490–440 Ma), the Tabberabberan (*c.* 50 million years, 430–380 Ma) and the Kanimblan (*c.* 60 million years, 380–320 Ma). All three cycles can be recognized to varying degrees in the North Queensland Orogen, with the Benambran and Tabberabberan cycles also seen in the New England Orogen. The early part of the Benambran cycle corresponds in time with the post-collisional phase of the Delamerian cycle.

Lachlan supercycle 1 (Benambran cycle)

Convergent phase. There are four key lithotectonic elements in the Benambran cycle (Figs 6, 7, 8): (1) widespread quartz-rich, craton-derived turbidites, overlain in the east by a condensed sequence of black shale; (2) intra-oceanic Macquarie Arc; (3) tholeiitic basalt-chert association with MORB-like chemistry; and (4) the ocean floor Narooma Terrane. An additional element is a very small outcrop of Ordovician limestone (Early Ordovician Digger Island Marlstone lying on Cambrian ?chert) on the Victorian south coast (VandenBerg *et al.* 2000).

(1) Early and Middle Ordovician turbidites in the Central and Eastern subprovinces contain both thick and thin beds of sandstone grading up into siltstones and interbedded with multiply-cleaved slates. VandenBerg *et al.* (2000) recognized several facies. Sandstones are generally quartz-rich with up to 10% detrital feldspar as well as detrital white mica, especially in the lower part of the sequence, which also includes lithic grains (M. Scott & O. Thomas, unpublished work). In the Tabberabbera area, basal sandstones are lithic and lie on Cambrian sedimentary rocks above Cambrian tholeiitic basalts (Crawford 1988; Fergusson 1998). In the Central and Eastern subprovinces, thin chert bands or lenses occur in the Early Ordovician in the Bendigonian (Stewart & Glen 1991) and the Chewtonian (Colquhoun *et al.* 2004). A prominent chert packet up to 100 m thick near the top of the turbidite packet is late Darriwilian, or Darriwilian–Gisbornian in age (Glen 1992; 1994; VandenBerg & Stewart 1992; Colquhoun *et al.* 2004*a*) (Fig. 8b). Turbidites, contourites or siltstones above the chert (Jones *et al.* 1993; Glen 1994; Thomas *et al.* 2002) pass up into a condensed sequence (*c.* 400 m thick) of black shale and siltstone (Bendoc Group, Warbisco Shale), which spans all or almost all of the Late Ordovician, beginning in the Gisbornian (Fig. 8b) (VandenBerg 1981; VandenBerg *et al.* 1992; VandenBerg & Stewart 1992; Glen 1992). In some localities, black shales pass up into buff-coloured shales and then local quartz sandstone (Glen & VandenBerg 1987; Thomas *et al.* 2002; Colquhoun *et al.* 2004). Another packet of Early and Middle Ordovician turbidites occurs in the Southwestern subprovince in central Victoria (Figs 4a, 6, 7, 8). Unlike the other turbidite packets, these contain abundant graptolitic shales, thin from west to east (VandenBerg *et al.* 2000) and lack cherts (Fig. 8b). Unlike the subprovinces to the east, there are Late Ordovician turbidites in this packet that pass eastwards into Late Ordovician black shales (VandenBerg & Stewart 1992).

Ordovician turbidites in northeastern Tasmania are dated poorly and consist of the sandstone-rich Tippogoree Group overlain by a slate-rich unit, containing a sole Middle Ordovician graptolite (Reed 2001). Reed (2001) suggested that these Ordovician units should be correlated with similar rocks of the Central subprovince (terminology of this paper) in the Tabberabbera area of central Victoria rather than with rocks of South-western subprovince as inferred previously (e.g. Powell *et al.* 1993*a*).

Comparison of the different turbidite packets is now possible with the introduction of the new conodont identification techniques of I. Stewart and I. Percival (e.g. Stewart 1988; Glen *et al.* 1990; VandenBerg & Stewart 1992; Percival *et al.* 2003). This comparison shows that rather than representing a homogeneous mud pile (Coney 1992), there are differences in facies, lithology and sequence between turbidite packets across the different subprovinces (summarized in Fig. 10b). These differences lead to the concept that the turbidites were deposited from different fan systems subsequently structurally juxtaposed along major faults. They are thus parts of different terranes (Glen *et al.* 1992; VandenBerg & Stewart 1992; Glen 1993; Glen & Percival 2002; 2003), rather than the one Bengal-size megafan (Cas 1983; Powell 1983; Coney 1992; Fergusson & Coney 1992a). This is discussed further in the last section of the paper.

A feature of all these turbidites is the pattern of detrital zircon ages with a dominance of 650 (600–500) Ma ages and a lesser population of 900–1200 Ma detrital zircons (Williams *et al.* 2002). Local differences between different terranes are emerging. The younger population constitutes the Pacific Gondwana population of Ireland *et al.* (1994). Veevers (2000*c*) suggested that the 600–550 Ma zircons came from the Beardmore–Ross Orogen area of Antarctica, with Goodge (2002) showing that the Ross Orogen began to emerge at 580–520 Ma.

Williams *et al.* (2002), on the other hand, suggested that the Mozambique Belt was the most likely source of the zircons.

(2) The intra-oceanic Macquarie Arc (Figs 4a, 6, 7, 8) existed from the earliest Ordovician to the Llandovery (494–438 Ma) and is now represented by four main structural belts of volcanic and volcaniclastic rocks, separated by younger Silurian–Devonian rifts (Glen *et al.* 1998). The Macquarie Arc grew in three pulses (Glen *et al.* 2003; Percival & Glen 2006) (Fig. 8b). Pulse 1, 490–475 Ma, is best developed in the western and central belts, where volcanic and volcaniclastic rocks pass up into siltstones. The early arc formed on a rifted fragment of Cambrian forearc crust, intruded by post-collisional volcanic rocks (Glen *et al.* 2003; Crawford *et al.* 2005) 1000 km east of the old Delamerian margin. It was growing actively at the same time as the Delamerian highlands were rising. After a *c.* 11 million years hiatus, possibly reflecting back-arc spreading, pulse 2 of arc growth occurred at 466–455 Ma. It is represented by volcanic and volcaniclastic rocks in the western and central belts passing into thinner-bedded turbidites to the east. After a partial hiatus in volcanism in the west of *c.* 4 million years, pulse 3 activity (*c.* 452–438 Ma) resulted in coherent volcanic rocks in the western belt and southern part of the central belt and *c.* 438 Ma (Llandovery) porphyries. In contrast to pulses 1 and 2, pulse 3 rocks are shoshonitic in chemistry (Glen *et al.* 2003; Crawford *et al.* 2005). This third pulse is coeval with Late Ordovician Alaskan-type zoned mafic to felsic igneous complexes intruding back-arc basin turbidites to the west (Suppel & Barron 1986; Elliott & Martin 1991).

Early workers maintained that there was interfingering between rocks of the Macquarie Arc and Ordovician craton-derived turbidites. However, there is no provenance mixing (Glen & Wyborn 1997; Colquhoun *et al.* 1999; Meffre *et al.* 2005) and there are faults between these coeval packages, which are thus separate terranes (Glen & Wyborn 1997; Glen *et al.* 1998; Glen & Percival 2003; Glen 2004; Meffre *et al.* 2005).

(3) An association of basalts, with MORB-like tholeiitic chemistry, and cherts occurs adjacent to major faults in the Lachlan Orogen – the Gilmore and Tullamore fault zones between the Eastern and Central subprovinces and along the Kiandra–Narromine Structure in the Eastern subprovince. This association has proven or inferred Ordovician ages (e.g. Basden 1990; Lyons & Percival 2002) and is inferred to represent igneous crust to the back-arc basin turbidites that was exhumed and translated along major faults.

(4) The Narooma Terrane (Figs 4b, 6, 8) is an Ordovician oceanic terrane that consists almost wholly of chert, but which coarsens at the top into argillite, siltstone, sandstone and conglomerate as it approached the Gondwana margin. This interpretation follows the model of Glen *et al.* (2004*a*) rather than those of Miller & Gray (1996) or Fergusson & Frikken (2003).

Elements of the Benambran cycle are also recognized from other orogenic belts. In the Delamerian Orogen, it is represented by platform deposits in west Tasmania (largely Gordon Limestone) and by shallow-water clastic rocks in the Koonenberry Belt (Figs 4, 7).

In the southern New England Orogen, the Benambran cycle is represented both west and east of the Peel–Manning Fault System (Glen & Scheibner 1993) (Figs 4b, 7). To the west in fault slices are arc-derived sediments, with blocks of Late Ordovician limestone (Cawood 1976), that lie unconformably on Early Ordovician and Cambrian strata (Cawood & Leitch 1985). The 'Trelawney Beds' of Philip (1966) are probably limestone olistoliths in Early Devonian units (P. A. Cawood cited in Furey-Greig 1999). In Queensland, Ordovician limestone occurs in the Devonian Calliope arc.

Benambran elements in accretionary complex rocks east of the Peel–Manning Fault System include:

(a) limestone blocks with Eastonian conodonts immediately east of the fault (Furey-Greig 1999). They probably represent seamount cappings (Scheibner 1973; Flood 1999), although Furey-Greig (1999) preferred a sedimentary derivation from volcanic islands and fringing reefs to the west;
(b) gabbro in the Peel–Manning Fault System at Attunga (locality of Attunga (at), Fig. 7) with zircons dated at 460 Ma (Watanabe *et al.* 1998) but corrected to 479 Ma by Fanning *et al.* (2002). Further south, phengite in a gabbro gave a K–Ar age of 481 Ma (Fukui *et al.* 1995);
(c) Ordovician blueschist and related rocks. At Port Macquarie (p, Fig. 7), white mica has yielded K–Ar ages of 444 Ma (M. A. Lanphere quoted by Scheibner 1985) and ages of 476 and 471 Ma (Fukui *et al.* 1995). White mica K–Ar ages of 482–467 Ma from blueschist knockers and related rocks (originally basalts and gabbros) in schistose serpentinite matrix along the Peel–Manning Fault System at Glenrock probably represent uplift ages: subduction

was, thus, earlier (Fukui *et al.* 1995; Offler 1999).

There are few data on the Benambran cycle in the Thomson Orogen. The generally low-grade metasediments encountered in drill holes in the west are undated and Murray (1994) suggested they are Cambrian to early Late Ordovician in age. Local metamorphic rocks have Cambrian to Early Ordovician K–Ar mica ages, and intrusive granites have Late Ordovician to Devonian K–Ar mica ages. On the eastern margin of the orogen (Fig. 7), the Late Ordovician Fork Lagoons beds (quartz-rich sandstones, conglomerates, limestones and felsic volcanic detritus and basalts with island-arc tholeiite affinity; Withnall *et al.* 1996) (Fig. 4c) are faulted against the Neoproterozoic–Cambrian rocks of the Anakie Inlier. Withnall *et al.* (1996) suggested correlations with arc-related Late Ordovician units in the North Queensland Orogen and it is possible the Fork Lagoons beds indicate extensions of this orogen to the south.

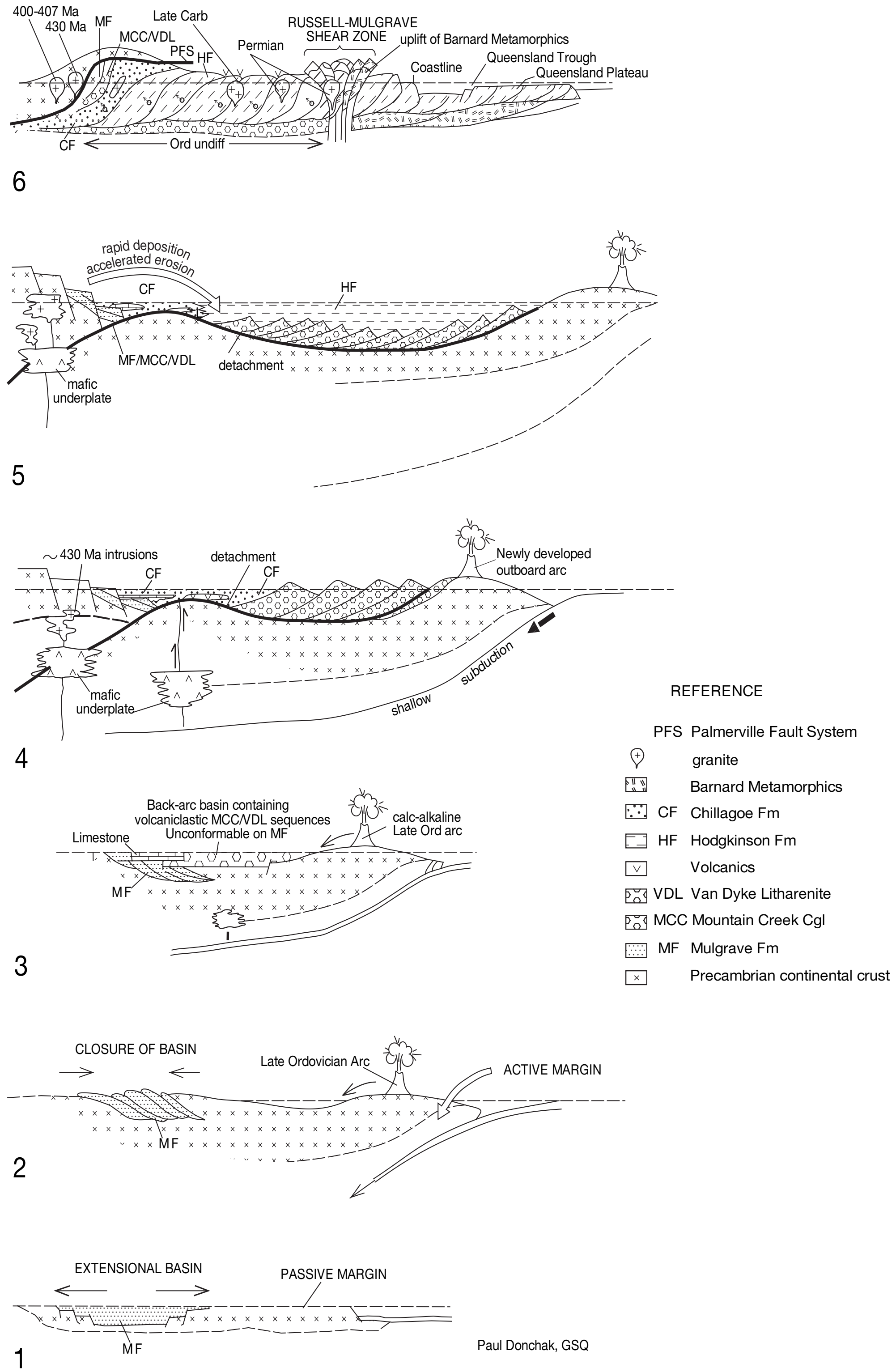
The Benambran cycle is also represented in the North Queensland Orogen, which contains evidence for Early and Late Ordovician subduction (Figs 4c, 7). The Lolworth–Ravenswood Block (Fig. 2) contains the *c.* 12 km thick Late Cambrian to Early Ordovician Seventy Mile Group, which shows a geochemical progression from intraplate basalt, andesite and alkaline magmas at the base to subduction-related, low- to medium-K calc-alkaline lavas higher up (Henderson 1986; Stolz 1995; Berry *et al.* 1992; Perkins *et al.* 1993; Paulick & McPhie 1999). Tectonically, this group represents initial rifting of a continental margin, leading to formation of a back-arc basin just inboard of a continental margin arc no longer preserved (Henderson 1986; Stolz 1995). Basement inliers were deformed strongly, with formation of a flat-lying extensional fabric at this time (Henderson *et al.* 2004).

This suprasubduction zone setting is consistent with the subduction signature of the coeval Late Cambrian to Middle Ordovician I-type, medium- to high-K granitoids in the Ravenswood Batholith (Hutton *et al.* 1994; Bain & Draper 1997). Model ages of 1120–1230 Ma, detrital zircons of 1100 Ma (Bain & Draper 1997; Draper *et al.* 1998) and large negative epsilon Nd indicate the involvement of old crust. Bain & Draper (1997) attributed the subduction signature to derivation from melting of underplated mafic melts.

The anomalous *c.* 150 km east–west trend of the Seventy Mile Group raises questions about the orientation of subduction – was it east–west (Henderson 1986), north–south (e.g. Kay 1998) or was it rotated from a SW–NE trend, parallel to the Diamantina River Lineament with subduction to the southwest (Stolz 1995)? Palaeomagnetic evidence in support of a 90° clockwise rotation between 425 Ma and 400 Ma was presented by McElhinny *et al.* (2003). This is consistent with the presence, to the present north, of rapidly deposited, moderately poorly sorted Ordovician quartz-rich turbidites derived from the craton to the west (Bain & Draper 1997). These turbidites constitute the Mulgrave Formation in the Hodgkinson subprovince (but contain more immature sandstones in the east) and the Early Ordovician Judea beds in the Broken River subprovince (Donchak 1993) (Figs 4c, 9(1)). Lenses of mafic volcanic rocks and jaspers in the turbidites and in slices along the Palmerville Fault System may be similar to the back-arc basin basalts in the Lachlan Orogen, although they also may relate to a suprasubduction zone environment (Bain & Draper 1997).

Middle to early Late Ordovician tectonism in the North Queensland Orogen in the Hodgkinson subprovince is reflected by deposition of Richmondian (*c.* =Bolindian) limestone above the Mulgrave Formation (Nicoll cited in Bain & Draper 1987) and beneath a conglomerate containing clasts of veined quartz-sandstone, derived from the Mulgrave Formation, and felsic-intermediate volcanics derived from an arc (Donchak 1993). A 455 Ma U–Pb age by C. M. Fanning on one dacite clast (cited by Donchak 1993; Bain & Draper 1997) suggests that these Late Ordovician sediments accumulated in a back-arc basin behind a west-dipping Late Ordovician subduction zone, and that a continental arc was established on a basement of 500 Ma metamorphic rocks (Fig. 9(3)). Initiation of subduction in the Middle Ordovician is approximately coeval with closure of the Early Ordovician sedimentary basin to the west (Fig. 9(2)).

Collisional phase. Evidence for some type of Middle Ordovician tectonism occurs in three orogenic belts. In the Lachlan Orogen, 455–458 Ma Ar–Ar mica ages were obtained from quartz veins in the Southwestern subprovince (Foster *et al.* 1999). In the southern New England Orogen, there is a mid-Ordovician unconformity (Cawood & Leitch 1985). In the North Queensland Orogen, Middle Ordovician deformation is recorded by an inferred unconformity above quartz-rich sediments of the Mulgrave Formation and below Late


400-407 Ma
430 Ma
MF
Late Carb
MCC/VDL
PFS
HF
Permian
RUSSELL-MULGRAVE SHEAR ZONE
uplift of Barnard Metamorphics
Coastline
Queensland Trough
Queensland Plateau
CF
Ord undiff
6
rapid deposition
accelerated erosion
CF
HF
MF/MCC/VDL
detachment
mafic underplate
5
~430 Ma intrusions
detachment
CF
CF
Newly developed outboard arc
mafic underplate
shallow
subduction
4
Back-arc basin containing volcaniclastic MCC/VDL sequences Unconformable on MF
Limestone
calc-alkaline Late Ord arc
MF
3
CLOSURE OF BASIN
Late Ordovician Arc
ACTIVE MARGIN
MF
2
EXTENSIONAL BASIN
PASSIVE MARGIN
MF
1
REFERENCE
PFS Palmerville Fault System
granite
Barnard Metamorphics
CF Chillagoe Fm
HF Hodgkinson Fm
Volcanics
VDL Van Dyke Litharenite
MCC Mountain Creek Cgl
MF Mulgrave Fm
Precambrian continental crust
Paul Donchak, GSQ

Ordovician arc-derived conglomerate in the Hodgkinson subprovince (Fig. 9(3)). This deformation is attributed to closure of the 'Mulgrave Basin' related to initiation of subduction to the east (Donchak 1993; Bain & Draper 1997). A major end Ordovician–Early Silurian deformation was responsible for thrusting between the craton and the western margin of the orogen (Bain & Draper 1997). In the southern part, it produced multiple deformation in the Mt Windsor Group, especially the east–west strike of the unit that reflects a structural position on the south-dipping limb of a D2 fold (Berry *et al.* 1992).

In the Lachlan Orogen, the Benambran cycle was terminated by the Benambran Orogeny that affected all of the orogen, except for the Melbourne Zone in the Southwestern subprovince. Ar–Ar mica ages indicate the Bendigo Zone underwent formation of east-vergent folds, thrusts and related cleavage around 440 Ma (Foster *et al.* 1999). In contrast, in the Central and Eastern subprovinces, the Benambran Orogeny consists of two identifiable phases phases over *c.* 10 million years (Packham 1969; VandenBerg 1999; Collins & Hobbs 2001; Glen *et al.* 2004*a*) (Fig. 4b). Phase 1 occurred around the end of the Ordovician (*c.* 443 Ma) and was marked by deformation of Ordovician turbidites and overlying black shales by folding, thrusting, major strike-slip faulting and multiple cleavage formation in response to east–west and north–south shortening. Part of the extinct Macquarie Arc was upthrust as it was accreted into back-arc basin turbidites (Glen *et al.* 2004*a*). Phase 1 was followed by extension leading to formation of deep- and shallow-water Llandovery sedimentary basins and is thus manifest by a major facies change from black shales to Llandovery turbidites (Glen *et al.* 2004*a*). Phase 2 (*c.* 433–430 Ma) of the Benambran Orogeny was marked by oblique thrusting of Ordovician turbidites in the Central subprovince over the Macquarie Arc, which was being translated to the southwest. Uplift of other parts of the Macquarie Arc, the folding and thrusting of Llandovery turbidites (VandenBerg 1999), and the syn-tectonic emplacement of some granites (VandenBerg 1999) are also parts of phase 2. In eastern Tasmania the Benambran Orogeny is poorly constrained in time, but involved recumbent folding with vergence to the northeast (Reed 2001).

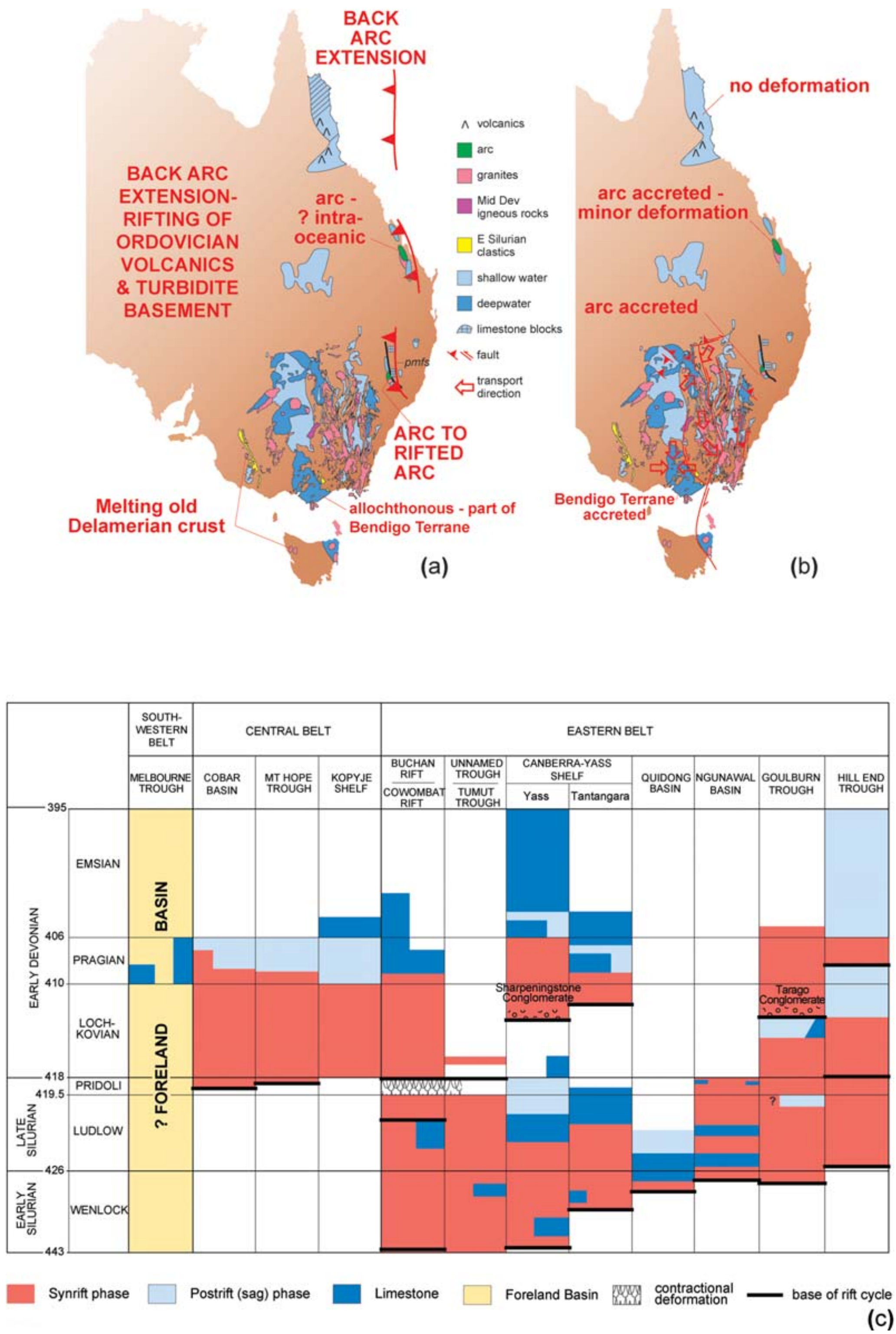
In the Koonenberry area of the Delamerian Orogen, the Benambran Orogeny is expressed by ESE-trending cleavage in post-Delamerian strata (Mills 2002). Mills also suggested that an ESE-trending fold belt in the SW corner of the Thomson Orogen formed during the Benambran Orogeny. In the New England Orogen, the Benambran Orogeny coincides with the hiatus between the Middle Ordovician Haedon Formation and Early Devonian strata at the base of the Tamworth Group (Cawood & Leitch 1985).

Lachlan supercycle 2 (Tabberabberan cycle)

Lachlan cycle 2 affected the Lachlan Orogen and the North Queensland Orogen, both of which have extensional character, lying behind the Gondwana-proto-Pacific plate margin identified from the New England Orogen by the presence of an intra-oceanic arc-subduction system. Granites of this cycle also intrude basement west and north of the North Queensland Orogen and the Delamerian Orogen west of the Lachlan Orogen back-arc basin elements

Convergent margin phase. Elements of a Late Silurian–Middle Devonian arc occur in fault-bounded blocks along the western flank of the Peel–Manning Fault System in the southern New England Orogen (Figs 4b, 10a). These volcanic rocks lie below Middle Devonian–Carboniferous strata of the Tamworth Trough, although they were placed originally in that forearc basin sequence by Crook (1960). Two volcanic packets are present: one Late Silurian to Middle Devonian and the other Middle to Late Devonian in age. The older packet of volcaniclastic, extrusive and intrusive rocks was interpreted as an intra-oceanic arc with a low-K calc-alkaline signature by Cawood & Flood (1989) and Offler & Gamble (2002). Both papers suggested the arc developed above a west-dipping subduction zone, although Aitchison & Flood (1995) argued for east-dipping subduction. Offler & Gamble (2002) suggested that the Middle–Late Devonian volcanic rocks represented inter-arc rifting (see also Stratford 1993; Aitchison & Flood 1995; Stratford & Aitchison 1996). The significance of a 436 Ma U–Pb zircon age from tonalite in the Pigna Barney Ophiolite Complex is uncertain*:* it could be related to pieces of an Ordovician arc or to

Fig. 9. Geological history of the central and northern parts of the North Queensland Orogen, Hodgkinson subprovince. From Paul Donchak, Geological Survey of Queensland, with permission.


BACK ARC EXTENSION
BACK ARC EXTENSION- RIFTING OF ORDOVICIAN VOLCANICS & TURBIDITE BASEMENT
arc - ? intra-oceanic
pmfs
ARC TO RIFTED ARC
Melting old Delamerian crust
allochthonous - part of Bendigo Terrane
volcanics
arc
granites
Mid Dev igneous rocks
E Silurian clastics
shallow water
deepwater
limestone blocks
fault
transport direction
(a)
no deformation
arc accreted - minor deformation
arc accreted
Bendigo Terrane accreted
(b)
SOUTH-WESTERN BELT
CENTRAL BELT
EASTERN BELT
MELBOURNE TROUGH
COBAR BASIN
MT HOPE TROUGH
KOPYJE SHELF
BUCHAN RIFT
COWOMBAT RIFT
UNNAMED TROUGH
TUMUT TROUGH
CANBERRA-YASS SHELF
Yass
Tantangara
QUIDONG BASIN
NGUNAWAL BASIN
GOULBURN TROUGH
HILL END TROUGH
395
406
410
418
419.5
426
443
EMSIAN
PRAGIAN
LOCH-KOVIAN
PRIDOLI
LUDLOW
WENLOCK
EARLY DEVONIAN
LATE SILURIAN
EARLY SILURIAN
BASIN
? FORELAND
Sharpeningstone Conglomerate
Tarago Conglomerate
?
Synrift phase
Postrift (sag) phase
Limestone
Foreland Basin
contractional deformation
base of rift cycle
(c)

Pb loss from the Cambrian arc (Kimbrough *et al.* 1993).

In the northern New England Orogen, Upper Silurian to Middle Devonian rocks within the area of the Yarrol Trough were placed in an island-arc setting by Marsden (1972) and grouped into the Calliope Volcanic Arc by Day *et al.* (1978) (Fig. 10a). These rocks are dominantly shallow-marine volcaniclastic sediments with varying amounts of felsic to mafic volcanic rocks. From geochemical data, Morand (1993) suggested that these rocks formed in a continental margin arc setting, whereas Offler & Gamble (2002) used REE data to argue for an island-arc setting. Messenger (1996) suggested a continental island arc undergoing local rifting. Murray (2003) suggested the 380 Ma Mt Morgan trondhjemite has an arc or rifted-arc geochemistry. In contrast, Bryan *et al.* (2003*b*) suggested formation in a back-arc basin and argued that zircon and Pb isotopic data support the involvement of continental crust in magma generation.

In the southern New England Orogen, the accretionary complex of the Tabberabberan cycle occurs as a narrow 10–20 km wide terrane east of the Peel–Manning Fault System (Fig. 10a). This is the Woolomin Terrane (Cawood & Leitch 1985), broadly correlating with the Djungati Terrane of Aitchison *et al.* (1992*a*), which comprises a fault-repeated basalt–chert–sandstone package (Woolomin Group) (Cawood 1982*b*; Aitchison *et al.* 1992*a*) (Fig. 11a). Radiolaria indicate a mid-Silurian age for basalt, Late Silurian–Frasnian age for chert and a Famennian age for siliceous chert and overlying volcanogenic sandstone (Ishiga *et al.* 1988; Aitchison *et al.* 1992*a*). Limestone olistoliths in the sandstone range in age from Late Ordovician through to the Devonian (Cawood 1980; Pickett 1982; Furey-Greig 1996; Flood 1999) and may represent partly accreted seamounts (Scheibner 1973; Aitchison *et al.* 1992*a*; Flood 1999).

In the northern New England Orogen, the presence of a Late Silurian–Middle Devonian conodont fauna (Fergusson *et al.* 1993) indicates some accretion in the Tabberabberan cycle, but accretion was mainly younger.

Back-arc system. In the Lachlan Orogen, the Tabberabberan cycle is characterized by major

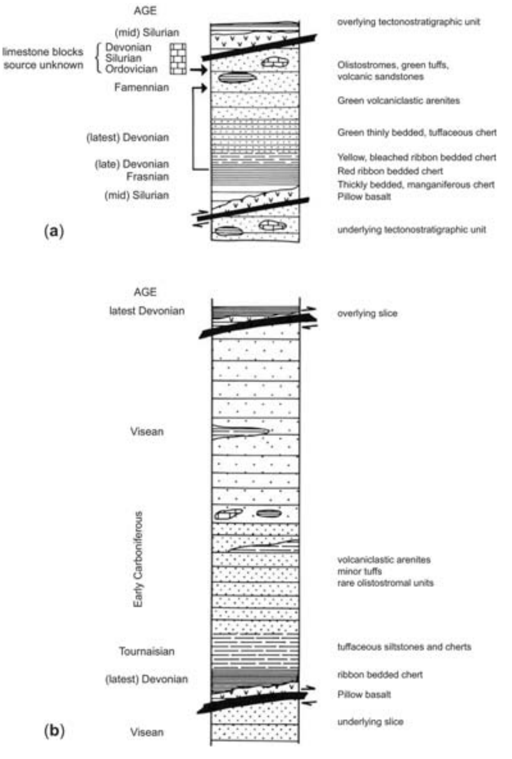


Fig. 11. Stratigraphy of two New England Orogen terranes: (**a**) Djungati terrane from the Tabberabberan cycle; (**b**) Anaiwan terrane in the Hunter–Bowen cycle. After Aitchison *et al.* (1992*a*) with permission of the journal of *Palaeogeography, Palaeoclimatology and Palaeoecology*, and the authors.

basin formation and emplacement of granitoids (Figs 4, 6, 10). Basins in the Eastern, Central and Western subprovinces formed by rifting or transtension. In the absence of published subsidence curves, three criteria suggest that basins in the Eastern, Central and Western subprovinces had rift and/or transtensional origins:

1. the presence of bounding and internal growth faults (e.g. Stuart-Smith 1990; Glen *et al.* 1996; Glen 1999; VandenBerg *et al.* 2000) revealed by thickness variations and/or facies changes (e. g., megabreccias

Fig. 10. Tabberabberan cycle. (**a**) Convergent margin and back-arc units. (**b**) Collisional phase showing complex thrust ± strike-slip faulting in the Lachlan Orogen with directions of maximum horizontal shortening indicated by open red arrows. (**c**) Simplified time–space plot of some Tabberabberan basins in the Lachlan Orogen.

on the margins of the Early Devonian Buchan Rift (VandenBerg *et al.* 2000);
2. the change in basin fill, from volcanic and/or coarse clastics at the base, passing up into limestone and shale, and/or fine-grained clastics, at the top. These changes suggest that basin growth may be divided into syn-rift and post-rift or sag phases (Fig. 10c) and, thus, that these basins formed as rifts, half-graben or transtensional basins (Glen 1992);
3. the rift geochemical signature of volcanic and volcaniclastic rocks, with most volcanic rocks plotting in the fields of an incipient back-arc, probable back-arc extension, continental rift setting, or even intra-arc settings (e.g. Collins 2002). Some volcanic units have subduction zone signatures indicative of melting of crustal or mantle material modified previously by subduction (Basden 1990; Dadd 1998; Watkins 1998*b*) (see also Part 3).

Using these criteria, it appears that most/all of the Middle Silurian–Middle Devonian basins in the Eastern and Central subprovinces formed during rifting that commenced around the lower part of the Wenlock. Multiple rift–sag sequences occur within some basins, indicating renewed extension in the middle to upper Lochkovian (Fig. 10b), leading to the formation of intra-basin unconformities below Lochkovian conglomerates from the Canberra–Yass Shelf and Goulburn Trough. Although attributed previously to the contractional Bowning Orogeny (Brown *et al.* 1968; Crook & Powell 1976; Owen & Wyborn 1979; Felton 1976; Bain *et al.* 1987), the unconformities are regarded here as extensional in nature.

It is suggested that the presence or absence of volcanic rocks in rift basins can be used to divide rifts into hot ones that contain felsic volcanic and volcaniclastic rocks with or without co-magmatic granites, and cold ones filled either by epiclastic or siliciclastic detritus (unpublished data). The Hill End Trough and Cobar Basin are the only two major cold rifts in the Central and Eastern subprovinces (Fig. 6) and both lie north of a major cross fault – the Lachlan Transverse Zone (Glen & Walshe 1999). This large-scale difference in heating of the extended crust is still not understood, but appears to be largely independent of basement type, with the Cobar Basin developed on a basement of deformed turbidites (Glen *et al.* 1996) and the Hill End Trough on a basement of Ordovician volcanic rocks (Glen *et al.* 2002; David *et al.* 2003)

To the south and west, the Southwestern subprovince was occupied by one large basin, the Melbourne Trough (now Melbourne Zone, Fig. 6), that was open to the east, and filled gradually from the west, south and southwest, right though to the late Early Devonian (Garratt 1983; VandenBerg *et al.* 2000; Powell *et al.* 2003), when detritus began to be supplied from an encroaching eastern source. In contrast to evidence of widespread extension in the three subprovinces above, there is little evidence of extension in the Southwestern subprovince. The Melbourne Trough is best understood as a composite foreland basin, yoked for some of its life to the deforming Bendigo structural zone to the west (western part of the Bendigo terrane) (Gray & Foster 1998; VandenBerg *et al.* 2000; Cayley *et al.* 2002) and containing packets of sandstone derived from the southwest and west as well as unknown areas to the southwest and south. In the Emsian, the Melbourne Trough was also yoked to the uplifted Tabberabbera Zone to the east, receiving volcanic-rich detritus from the east (Powell *et al.* 2003). Despite this foreland basin model, some extension in the eastern sector is reflected in part by the emplacement of the Middle Devonian (375 Ma) Woods Point Dykes with shoshonitic chemistry (Bierlein *et al.* 2001).

In eastern Tasmania, the Tabberabberan cycle is represented by turbidites of the Ludlow to Pragian Panama Group, intruded by Early Devonian I- and S-type granitoids (Reed 2001).

In the western Tasmanian part of the Delamerian Orogen, the absence of a Benambran Orogeny means the stable Ordovician platform (Gordon Group) continued into the Late Silurian, with shelf deposition of the Eldon Group (Seymour & Calver 1995). Reed (2001) pointed out that the overlying deep-water Early Devonian turbidites were correlatives of the Panama Group in eastern Tasmania, thereby indicating that the two terranes were close to each other by then.

In the western Victoria part of the Delamerian Orogen, the Tabberabberan cycle is represented by deposition of fluviatile and marginal marine sandstone-rich sediments of the Grampians Group, which are Late Silurian in the upper part (VandenBerg *et al.* 2000) (Figs 6, 13a). This unit was then thrust imbricated, extended (420–410 Ma) and, subsequently, intruded by I- and A-type Early Devonian 410–400 Ma post-tectonic granites (Cayley & Taylor 1996). These were then overlain by the Rocklands Rhyolite dated at 410 Ma (VandenBerg *et al.* 2000).

In the Koonenberry area of the Delamerian Orogen, the Tabberabberan cycle is represented by deposition of the Late Silurian–Devonian Mt Daubney Formation (Fig. 6), followed by its deformation in the late Early Devonian (Neef *et al.* 1989; Buckley 2003).

In the Thomson Orogen, the Tabberabberan cycle is represented by development of the Adavale Basin and adjacent troughs that are concealed beneath Permo-Triassic basins and younger cover (Figs 10a, b). Early Devonian felsic volcanic and volcaniclastic rocks pass up with apparent conformity into Middle Devonian marine sediments and evaporites (Murray 1994).

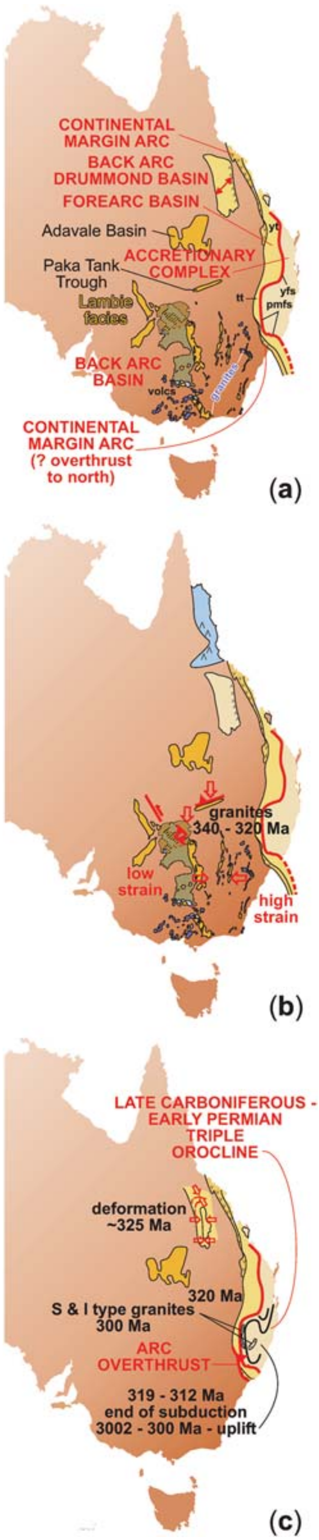
In the North Queensland Orogen, there is no major mid-Devonian deformation: deformation in the Broken River and Hodgkinson subprovinces occurred later, around the Devonian–Carboniferous boundary and, as a result, the Tabberabberan cycle extends into the early Tournaisian. There are, however, low-angle unconformities and disconformities around the Middle–Late Devonian boundary in both the Lolworth–Ravensworth and Broken River subprovinces (Bain & Draper 1997). The Tabberabberan cycle was marked by the formation of rift basins, filled by limestone, turbidites and shallow to continental siliciclastic rocks, and by the emplacement of granites. These rocks are mainly overlain by Viséan volcanic rocks deposited in graben (Bain & Draper 1997).

In the Hodgkinson subprovince, the Tabberabberan cycle began with extensional faulting that created accommodation for deposition of thin-bedded siliciclastic turbidites in the lower part of the Chillagoe Formation (Fordham 1993) and early to mid-Llandovery submarine volcaniclastic rocks and basaltic lavas (of possible MORB affinity), especially in the north. By the middle to late Llandovery, carbonate platforms developed in shallow sub-basins. Coeval deeper sub-basins received turbidites and continent-derived sediments. Carbonate deposition generally ceased in the Early Devonian (around 410 Ma, in the late Lochkovian), with the collapse of the carbonate platform in a renewed phase of basin extension, characterized by deposition of craton-derived breccia fans and conglomerates in submarine channels (Fordham 1993). This event was synchronous with the emplacement of the Cape York Peninsula Batholith in basement rocks to the west (Fig. 2) (Domagala *et al.* 1998). Conglomerates were succeeded by quartz-intermediate turbidites of the Hodgkinson Formation that covered a large area (Fig. 9(5)) and which are at least as young as mid-Famennian (Domagala *et al.* 1998). Most of these turbidites were derived from Proterozoic basement and >420 Ma Silurian–Devonian granites, although some sandstones in the east contain detritus from the old Ordovician arc (Domagala *et al.* 1998). The presence of lenses of submarine basalt, chert and local limestone and broken formation has been used in the past to invoke an accretionary prism setting for the Hodgkinson Formation (Henderson 1987), but the back-arc basin rift model of Fawckner (in Arnold & Fawckner 1980), Donchak (1993) and Domagala *et al.* (1998) is followed in this paper.

The Broken River subprovince to the southwest is dominated by siliciclastic rocks, although there is a local regression leading to formation of carbonates in the Emsian–Givetian. These pass up across a low-angle unconformity to disconformity into shallow-water to alluvial strata in the pull-apart Bundock Creek Basin. In the Lolworth–Ravenswood Block, Givetian arkoses, limestone and sandstone are overlain with local unconformity by Frasnian–Famennian continental sediments that pass up into alternating shallow-marine to continental sediments (fill of the Burdekin Basin) below a late Tournaisian unconformity and younger felsic volcanic rocks.

Tabberabberan cycle granites. Tabberabberan cycle granites occur in North Queensland and in the Lachlan Orogen. In North Queensland, granites were emplaced into the North Queensland Orogen and also into Precambrian basement to the west (e.g. the *c.* 430 Ma Nundah Granite, Fig. 9) (Bultitude *et al.* 1993). Basement northwest of the North Queensland Orogen was intruded by the >450 km long Cape York Peninsula Batholith, which consists of a core of I-type granites surrounded by dominant S-type granites that are mainly Early Devonian in age. Both types have Early Proterozoic model ages, suggesting derivation from underlying old continental crust. Within the North Queensland Orogen, Tabberabberan cycle granites in the Ravensworth Batholith in the Lolworth–Ravensworth Block comprise regional aureole I- and S-type granites (the oldest, 435–424 Ma), followed by 425–410 Ma oxidized I-type granites, 410–400 Ma regional aureole granites and 400–380 Ma high-level granites. The oxidized I-type granites have subduction-related chemistry, but were believed by Draper & Bain (1997) to have been derived from mid–deep crustal sources as a result of underplating by mafic magmas during regional extension.

In the Lachlan Orogen, Tabberabberan cycle granites are prevalent in the Eastern and Central subprovinces where they occupy up to



36% of the surface area (Chappell *et al.* 1988) (Fig. 6). Granites are divided into I-type granites in the east and mixed I- and S-type granites in the west (Chappell & White 1974): very few granites have mixed or zoned I–S character. I-types are subdivided further into high and low temperature types and fractionated or non-fractionated types (Chappell *et al.* 1998; 2000).

Some granites were emplaced into the extending upper crust, where they are co-magmatic with, and overlain by, Late Silurian and Early Devonian felsic volcanic rocks erupted in extensional basins (Wyborn & Chappell 1986). Other, foliated, granites were emplaced into the middle crust at depths of *c.* 10 km and then exhumed by thrusting a further 1–2 km (V. Morand, cited in Glen & Wyborn 1997).

Granite emplacement ages migrate eastwards from the Silurian in the Central subprovince into the Silurian to Devonian in the Eastern subprovince (e.g. Powell 1983, 1984*a*; Collins 2002), but there are young Late Devonian–Carboniferous granites in central Victoria in the Southwestern subprovince as well in the eastern part of the Eastern subprovince. Some of this 'migration' might actually reflect cooling patterns, since recent zircon dating has shown that granites thought previously to be Late Silurian to Early Devonian from K–Ar and Rb–Sr dating are older and formed during early parts of the Tabberabberan cycle, if not at the end of the Benambran cycle, e.g. Cooma granodiorite (Williams 2001). If the model of Glazner *et al.* (2004) applies to Lachlan granites, individual plutons may have been emplaced over millions of years, and can no longer be represented by a single zircon date.

Collisional phase. In the southern New England Orogen, Flood & Aitchison (1992)

Fig. 12. Kanimblan cycle and Hunter–Bowen supercycle. (**a**) Fluviatile Lambie facies represented by mustard-coloured units in the Lachlan, Delamerian and Thomson orogens. Hachured area in the Lachlan Orogen represents area of concealed Lambie facies developed above Tabberabberan cycle strata. Elements of Hunter–Bowen supercycle also shown: back-arc Drummond Basin, continental margin arc, forearc basin (tt, Tamworth Trough; yt, Yarrol Trough) and subduction complex and accreted terranes outboard of major faults (pmfs, Peel–Manning Fault system; yfs, Yarrol Fault System). (**b**) Collisional and post-collisional phase, Kanimblan cycle.(**c**) Hunter–Bowen supercycle, deformation phase, cycle 2.

suggested that the intra-oceanic arc of the Tabberabberan cycle became accreted to the Australian plate in the Late Devonian (Fig. 10b). This is controversial, however, and no clear evidence of contractional deformation related to arc accretion has been identified. The candidate Bective unconformity below the Late Devonian Keepit Conglomerate is a semi-regional disconformity only, and is not present in the centre of the Tamworth Trough (Russell 1979).

In the northern New England Orogen, proponents of the intra-oceanic model for the Calliope arc argue that an unconformity close to the Middle–Late Devonian boundary reflects accretion of the arc to the Gondwana margin (e.g. Murray *et al.* 2003). Others (Leitch *et al.* 1992; Morand 1993) have argued that the unconformity is low angle, deformation was minor and that there was no break at the base of the overlying sequence (see also Bryan *et al.* 2003); C. Murray (pers. comm. 2004) points to removal of a major stratigraphic unit along the unconformity.

In the Lachlan Orogen, major basin inversion occurred in the late Early–Middle Devonian, with earlier Devonian deformations reflecting localized strike-slip tectonics. There are differing views as to whether these deformations are part of a longer-lived deformation prograding from the west (Gray & Foster 1997; Gray *et al.* 1997), or are related to the oblique strike-slip collision of an allochthonous terrane into the Southwestern subprovince (Glen *et al.* 1992; VandenBerg & Stewart 1992; Willman *et al.* 2002; see Part 3).

Middle Devonian orogeny is reflected by basin inversion, with growth faults undergoing varying degrees of reverse/oblique reactivation and development of thin-skinned and basin-inversion structures (e.g. Glen 1995) (Fig. 10b). In the Eastern subprovince, Middle Devonian basin inversion involved fold and cleavage formation and NE–SW thin-skinned thrusting in the Hill End Trough (Vassallo *et al.* 2003), deformation of the shelf to the west (*c.* 380 Ma cooling ages, Glen *et al.* 1999) and in the 'Captains Flat Trough' to the south (370–380 Ma K–Ar ages on sericite, 374 Ma Rb–Sr age on biotite, Abell 1993). The Jemalong Trough (Fig. 6) (Raymond *et al.* 2001) and Cowra Trough were deformed at *c.* 393 Ma (Raymond *et al.* 2001), while the overlying Middle Devonian A-type volcanic rocks of the Rocky Ponds Group (Dulladerry Volcanics) were deformed *c.* 372 Ma (Raymond *et al.* 2001). The presence of metamorphic biotite in deformed sedimentary rocks of the Hill End Trough (Vernon & Flood 1979) and 'Captains Flat Trough' (Smith 1969; Barron 1999; Abell 1993) and metamorphic white mica in the eastern part of the Cobar Basin (Brill 1989; Glen *et al.* 1992) implies more intense deformation and structural thickening than first apparent (e.g. Vassallo *et al.* 2003).

The N–S elongate shapes of some Silurian–Devonian granites (Fig. 6) also reflect this Middle Devonian deformation, with the formation of solid-state foliations and well developed S–C fabrics and mylonite zones on the (commonly) eastern margins of mid-crustal granites such as the Wyangala Batholith (Morand 1988; Paterson *et al.* 1990; Lennox *et al.* 1998) and Wologorong Batholith (Vernon *et al.* 1983). Cooling ages of *c.* 380 Ma (Foster *et al.* 1999; Glen *et al.* 1999) have been obtained from Ar–Ar dating of metamorphic biotite. The eastern, highly strained, margin of the Murrumbidgee Batholith might also have been deformed at this time. No evidence is found for the concept that rising granites were responsible for deformation seen in Silurian–Devonian rifts through downward movement (or sagductional deformation) of upper crustal rocks lying between plutons or batholiths (Warren & Ellis 1996; Blevin 1998).

The last stage of this Tabberabberan Orogeny in the Eastern subprovince was marked by brittle conjugate NE- and NW-trending faults offsetting major plutons. These faults produced small amounts of extension parallel to the orogen and were, thus, able to distribute the decreasing amounts of east–west shortening in a crust strengthened by cooling granites (Lambert & White 1965; Glen 1992).

An earlier deformation in the Early Devonian is concentrated along the boundary between the Eastern and Central subprovinces and reflects thrusting and translation of the Central subprovince to the SSE (Glen 1991; Morand & Gray 1991). Deformation was concentrated along a linked system of major and splay faults – the Tullamore Fault Zone, Gilmore Fault Zone and Indi–Long Plain Fault – that marks the boundary between these subprovinces (Fig. 6). To the east, the Tumut Trough was inverted between 418 Ma and 411 Ma (Basden 1990) and the largely transtensional rifts of the Cowombat Trough in Victoria were deformed by the Bindian deformation around 418 Ma (VandenBerg *et al.* 2000). This Bindian deformation is also recorded by Ar–Ar dates of 409 Ma on cleavage sericite at Peak Hill (Perkins *et al.* 1995) and 411 Ma on sericite in the Booberoi Fault Zone (dating by Foster *et al.* 1999, cited by Lyons 2000). To the west, the eastern margin of the Cobar Basin was deformed at 400–395 Ma, according to whole-rock Ar–Ar dating (Glen *et al.* 1992*b*). Movement on the Indi Fault is dated at 405 Ma (Foster *et al.* 1999) while faults near the leading edge of the south-moving Central subprovince have Ar–Ar mica ages of 413–395 Ma (Foster *et al.* 1999). The latitudinal boundary with the Thomson Orogen underwent major dextral strike-slip (Glen *et al.* 1996) probably after contractional dip-slip faulting.

Interpretation of geophysical data, coupled with limited outcrop information, suggests that significant basin inversion also occurred in the Western subprovince in the mid-Devonian (Glen *et al.* 1996).

In the Southwestern subprovince, most Middle Devonian structures formed in response to broadly east–west shortening coupled with lesser north–south shortening. Effects of both shortenings are best seen in the Melbourne Zone, where the low-strain, western part contains curvilinear (NE)–NW-trending folds, refolded about east–west folds (see summary in VandenBerg *et al.* 2000). East–west strain becomes higher eastwards, with the eastern Melbourne Zone containing west-dipping thrusts (Gray 1995) and duplexes in Cambrian igneous units (VandenBerg *et al.* 1995). Ar–Ar dating of mica gives ages of 415–390 Ma (Foster *et al.* 1999) (Fig. 6).

A two-stage Tabberabberan deformation is present in eastern Tasmania. Early Devonian, pre-granite, NE-vergent structures were followed in the late Early to Middle Devonian by syn-granite structures as this part of the Lachlan Orogen was thrust southwest over western Tasmania (Reed 2001). As a result of this deformation, the Delamerian Orogen in western Tasmania underwent tightening of older Delamerian folds, generation of NW-trending folds, cleavage and faults, and heating that led to the intrusion of syn- to post-tectonic granites (Berry 1994).

No clear Tabberabberan deformation has been recognized from the concealed Adavale Basin, although Evans *et al.* (1990) suggested the basin was converted to a foreland basin around this time.

In the North Queensland Orogen, the Tabberabberan cycle was brought to a close by an Early Carboniferous deformation (Fig. 9(6)). In the Hodgkinson subprovince, this is best constrained by a syn-tectonic 357 Ma granite that was intruded during formation of a widespread S2 foliation (Davis *et al.* 1998). In the Broken River subprovince, deformation separates little-deformed Tournaisian sedimentary rocks in the Clarke River Basin from underlying strata (Zucchetto *et al.* 1999) (Fig. 4c). This deformation did not apparently affect the Lolworth–Ravenswood subprovince to the southwest, where sedimentation in the Burdekin Basin continued into the Tournaisian.

Lachlan supercycle 3 (Kanimblan cycle)

Elements of the Kanimblan cycle occur in the Lachlan and Delamerian orogens (Figs 4, 12a).

Rifting and loading phases. In the Eastern and Central subprovinces of the Lachlan Orogen, a brief late Early–Middle Devonian rifting event marked the onset of the Kanimblan cycle and generated restricted A-type volcanic rocks and local granites and rift-related sediments (McIlveen 1974; Dadd 1992; Wyborn & Owen 1986; Raymond 1996; Collins 2002) (Figs 4, 6). In some areas there is a low-angle unconformity beneath 3–4 km of largely fluviatile sandstone, conglomerate and siltstones/mudstones that were deposited in one or more terrestrial basins and which constitute the Lambie (or Lambian) facies of Powell (1984*b*) (Fig. 12a). However, in general, rocks of the Lambie facies lie on a range of older stratigraphic units that were assembled in the preceding Tabberabberan Orogeny (Glen & Watkins 1999). In the Western, Central and Eastern subprovinces, rocks of the Lambie facies either pass up from a basal marine interval into continental fluviatile deposits, or contain a marine interval (that disappears westwards) near their base. Further west, in the Central and Western subprovinces, the Mulga Downs Group (Glen *et al.* 1996) contains, in its lower part, a late Early–Middle Devonian package (Bembrick 1997).

Rocks of the Lambie facies are overwhelmingly quartz-rich and possibly derived from major uplift in central Australia (Alice Springs Orogeny), although there are local facies variants, especially near the base and variations in palaeocurrent directions. One prominent interval with volcanic detritus was derived from the south (Powell 1984*b*), another from the arc in the New England Orogen to the north (Powell *et al.* 1984). It is still uncertain whether the Lambie facies formed a continuous 3–4 km thick blanket across the Lachlan Orogen or was deposited in interconnected intermontane basins (Powell 1984*b*). O'Halloran & Cas (1995) favoured intermontane basins in central Victoria and recognized intra-formational unconformities that they attributed to syn-depositional contractional deformation, probably reflecting long-lived movement on the

Governor Fault Zone between the Eastern and Central subprovinces (Fig. 6).

Different relations occur in the Southwestern subprovince. Here, the Kanimblan cycle began with the widespread intrusion of high-level S- and I-type granites and the outpouring of associated felsic volcanic rocks preserved in cauldron complexes of the Central Victorian Magmatic Province (Figs 4a, 6, 12a). Local mafic magmas are shoshonitic, characteristic of post-orogenic regions (VandenBerg *et al.* 2000). The succeeding Lambie facies, while broadly similar to units further east, contains a prominent volcanic interval marked by rhyolitic ignimbrites derived from cauldron activity and is intruded by high-level granites (VandenBerg *et al.* 2000).

Lambie facies rocks in the Delamerian Orogen are restricted to the Koonenberry area, where a late Early–Middle Devonian unit and a Late Devonian unit are preserved in the largely subsurface Bancannia and Menindee troughs as well in the Koonenberry Belt (Sharp & Buckley 2003; Neef 2004) (Figs 6, 12a).

In the Thomson Orogen, Late Devonian red-beds in the upper part of the concealed Adavale Basin correspond to rocks of the Lambie facies (Murray 1994). The concealed E–W- to WSW-trending Paka Tank Trough in the southern part of the orogen is of this age as well (Alder 1999) (Fig. 12a).

In the North Queensland Orogen, Lambie facies rocks are restricted to a *c.* 600 m thick sequence of siltstones and tuffaceous sandstone in the Lolworth–Ravensworth Block and in the Bundock Basin in the Broken River subprovince (Bain & Draper 1997).

Deformation phase. The Kanimblan Orogeny, the last regional deformation to affect the Lachlan Orogen, is dated at *c.* 340 Ma. East–west Carboniferous shortening (with lesser strike-slip movement) was intense in the Eastern subprovince and led to out-of-sequence thrusting (Glen, unpublished data), with formation of high- and low-strain zones that are best recognized by deformation of the Late Devonian Lambie facies rocks. High-strain areas are reflected by formation of map-scale, narrow, elongate zones of Late Devonian strata (Fig. 12b) that occupy the cores of tight synclines (Powell 1984*b*), commonly occupying footwalls of major N–S thrusts (Glen 1992). N–S D2 folds, faults and cleavage in the Hill End Trough are also Carboniferous in age and reflect a second phase of basin inversion by thick-skinned reverse faults (Vassallo *et al.* 2003). This intense Carboniferous deformation and regional cleavage-forming event in the NE of the Eastern subprovince (dated at *c.* 340 Ma) was followed by granite emplacement (Pogson & Watkins 1998).

Carboniferous strain decreases to the south and west, where Lambie facies rocks have been deformed into large N- to NW-trending broad synclinoria and anticlinoria that may also show effects of north–south shortening (Figs 6, 12b) (Glen 1992; Glen *et al.* 1996). Effects of the Kanimblan Orogeny extend westward into the Delamerian Orogen: in the Koonenberry area, the Bancannia Trough was inverted and there was sinistral reverse movement on the NW-trending Koonenberry Fault that produced folding and thrusting in the Lambie facies rocks in the footwall (Sharp & Buckley 2003; Neef 2004).

In the Thomson Orogen, deposition in the Adavale Basin was terminated by deformation in the mid-Carboniferous (Murray 1994). North–south shortening across the suture between the Lachlan and Thomson orogens resulted in basement rocks thrust south over the northern margin of the Paka Tank Trough (Fig. 12b) and in inversion of the trough itself (Alder 1999).

Post-collisional phase. In the Eastern subprovince of the Lachlan Orogen New South Wales, the Kanimblan Orogeny was followed by intrusion of high-level I-type granitoids (Pogson & Watkins 1998; Shaw & Flood 1993) generated by melting of underlying Ordovician volcanic rocks (Watkins 1998*a*) (Fig. 6).

Hunter–Bowen supercycle

The Hunter–Bowen supercycle records the Middle Devonian to Triassic convergent margin development of East Gondwana that is expressed by the evolution of the New England Orogen and the Bowen–Gunnedah–Sydney basin system to the west. In the New England Orogen, this supercycle is built on earlier elements of the Lachlan and Tabberabberan cycles.

From west to east, elements of this classical margin orogen include an arc, a forearc basin and subduction complexes together with accreted terranes (Leitch 1974; Day *et al.* 1978; Cawood 1982; Cawood & Leitch 1985; Leitch & Scheibner 1987; Scheibner 1989; Aitchison & Flood 1992) (Figs 12a, 13a). In this paper, this Devonian–Triassic history is divided into four cycles, but the changeover between cycles 1 and 2 is not clear-cut everywhere and further work may result in the two cycles being combined.

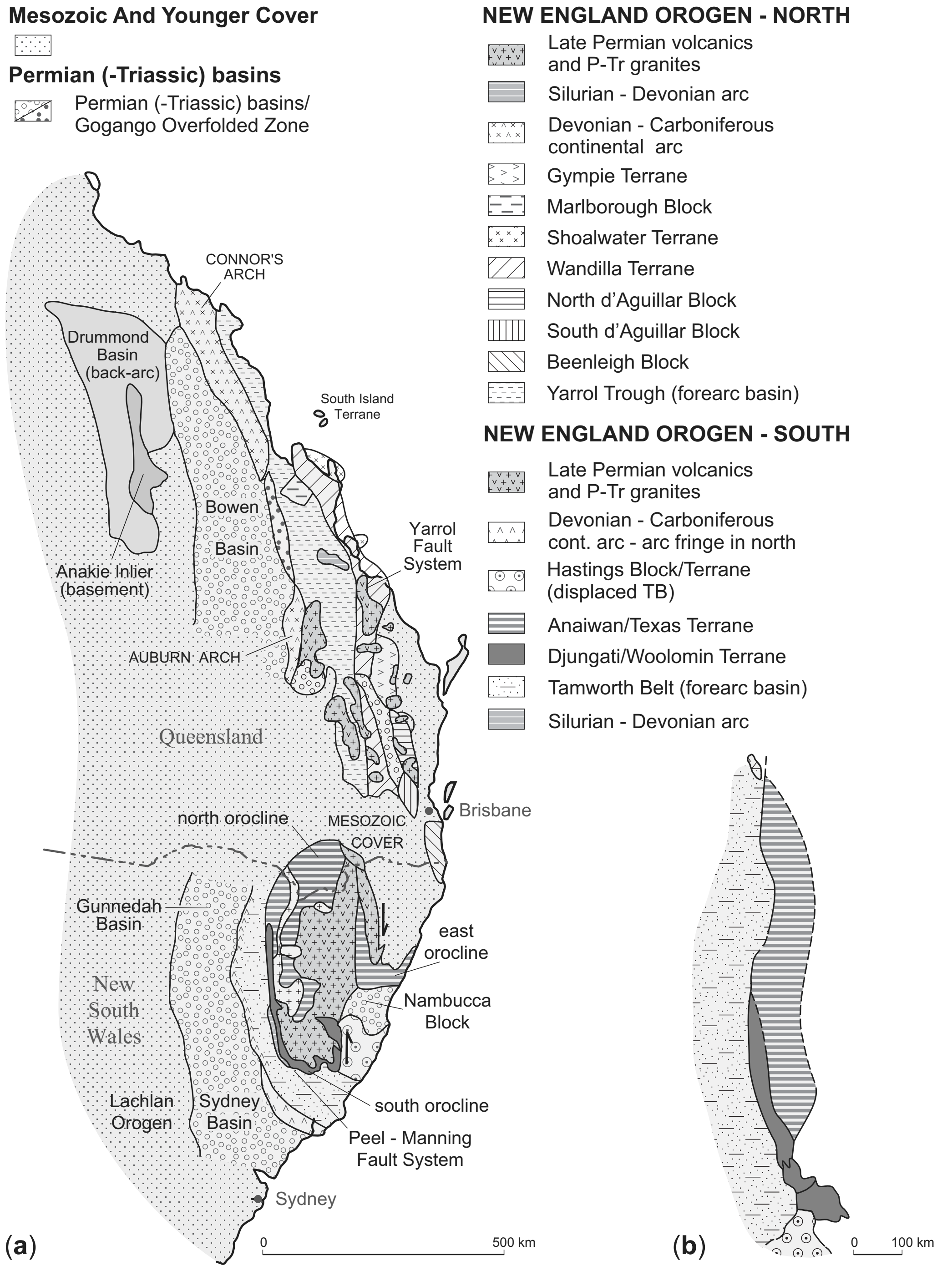


Fig. 13. Map of relations in the New England Orogen: (**a**) deformed state geology (note the triple orocline); (**b**) undeformed New England Orogen, after Cawood & Leitch (1985). Late Carboniferous–Early Permian palaeogeography derived from restoring the orocline.

- Cycle 1 (Late Devonian): characterized by an east-facing Late Devonian continental or intra-oceanic arc dominated by intermediate volcanism, a forearc basin (Yarrol Trough in north, Tamworth Trough in south) bounded to the east by the Yarrol Fault and Peel–Manning Fault System, and subduction complexes with accreted terranes. A possible back-arc basin occurs in the north (Drummond Basin) but not in the south (Fig. 12a).
- Cycle 2 (Carboniferous): characterized by an east-facing Carboniferous Andean-type arc dominated by felsic ignimbrites and granites in the west, passing eastwards into the Carboniferous part of the Yarrol and Tamworth troughs, and then into Carboniferous subduction complexes and accreted terranes.
- Cycle 3 (Early Permian): characterized by crustal extension associated with inception of the Bowen–Gunnedah–Sydney basin system and initial phase of the Hunter–Bowen Orogeny.
- Cycle 4 (Permian to Triassic): characterized by arc magmatism, the foreland basin stage of development of the Bowen–Gunnedah–Sydney basin system, widespread felsic igneous activity in the North Queensland Orogen to the west and terminated by the main phase of the Hunter–Bowen Orogeny, which reflects accretion of an intra-oceanic arc.

Veevers (2000*d*) has discussed cycles 3 and 4 in detail which he subdivided into seven stages.

Cycle 1: Late Devonian convergence

In Queensland, the rhyolitic to basaltic Drummond Basin, developed on crust of the Thomson Orogen, is thought to be a back-arc basin to the New England Orogen (Day *et al.* 1978; Murray 1986) (Fig. 4c). Note that the partly coeval development of the North Queensland Orogen to the north is interpreted to be part of the Tabberabberan cycle (Fig. 4c). This basin was initiated in the Late Devonian (Famennian) in the north but the Carboniferous (Tournaisian) in the south (Henderson *et al.* 1998), by NE–SW extension (Johnson & Henderson 1991) (Figs 15a, 16a). Volcanism and granite intrusion in the Anakie Inlier further west reflect coeval heating of the Thomson crust (Henderson *et al.* 1998) (Fig. 4c). Wood & Lister (2004) suggested that the two areas were linked as lower and upper plates of an extensional system.

The Late Devonian arc in the northern New England Orogen is best represented by the Campwyn Volcanics that were deposited on its eastern flank (Fergusson *et al.* 1994), although Bryan *et al.* (2003*a*) have suggested these rocks are rift related.

In the southern New England Orogen, an arc along the inboard part of the orogen is not preserved (Fig. 12c), but is inferred from large Late Devonian olistostromal blocks of andesitic volcanic rocks in the inboard part of the forearc basin, the Tamworth Trough (Brown 1987). This Baldwin Arc developed on oceanic or thin continental crust (Mollis 1988). The forearc basin to the east, the Tamworth Trough, was filled by marine strata that become finer grained and of deeper-water character eastwards, towards the Peel–Manning Fault System that marks the preserved outboard edge of the basin. Late Devonian trough-fill consists of mudstone packets that alternate with packets of sandstone and conglomerate that thin to the east.

In the northern New England Orogen, the Yarrol Trough is the equivalent forearc basin (Figs 12a, 13a). Late Devonian strata consist of volcaniclastic sandstone and conglomerate, derived from the andesitic arc to the west, interbedded with sediments and some limestone (Yarrol Project Team 1997; Fordham *et al.* 1998). Minor andesitic lavas, dacites and rhyolitic ignimbrites are present.

Bryan *et al.* (2001) suggested recently that the Yarrol Trough formed as a back-arc basin to an arc to the east, in an area now occupied by accreted terranes. While this model is based on arguments that volcanic rocks in the trough have a back-arc basin geochemical character and were derived from west- rather than east-flowing currents, on a more regional scale it implies a major decoupling between the northern and southern parts of the New England Orogen. Murray *et al.* (2003) rejected this back-arc model. Bryan *et al.* (2003*b*) re-affirmed their back-arc interpretation and Leitch *et al.* (2003) supported the forearc interpretation.

East of the Peel–Manning Fault System in the southern New England Orogen, Silurian–Early Devonian subduction continued into the Late Devonian without interruption. As a result, the upper parts of thrust slices (Fig. 11a) contain Famennian (Late Devonian) volcanolithic sandstone beds (with olistoliths), suggesting deposition not too remote from the subduction-related arc (Aitchison *et al.* 1992*a*). Ultramafic units of tholeiitic affinity along, and east of, the Yarras Fault were interpreted as a fragment of an accreted Middle to Late Devonian (374 Ma) arc by Aitchison *et al.* (1994).

In the northern New England Orogen, Late Devonian conodonts in the Beenleigh terrane indicate subduction in the Late Devonian (Aitchison 1988). Basalt dykes and fault-bounded blocks in the Neoproterozoic Marlborough Ophiolite have a Middle Devonian Sm–Nd isochron (380 Ma) and trace element data suggestive of an intra-oceanic island arc (Bruce & Niu 2000*b*). These relations suggest that the arc was probably built on a Neoproterozoic oceanic crust, lay only a short distance offshore (C. Murray & P. Blake unpublished geochemical data) and was accreted to Gondwana by the Early Permian (Bruce & Niu 2000*b*).

Cycle 2: Carboniferous convergence

In Queensland, the back-arc Drummond Basin continued to evolve, receiving basement-derived, Early Viséan quartz-rich detritus (Stage 2 of Johnson & Henderson 1991; Henderson *et al.* 1998) followed by Mid–Late Viséan ?arc-derived volcaniclastic rocks (Stage 3 of Johnson & Henderson 1991) (Fig. 4c).

In the northern New England Orogen, the Carboniferous continental margin arc is represented by granitic and mafic to silicic rocks of the Connors and Auburn arches arc (Day *et al.* 1978, but not Bryan *et al.* 2003*b*). These two arches lie along-strike, separated by the Gogango Overfolded Zone (Fig. 13a), which represents a part of the Permian Bowen Basin succession strongly deformed and thrust westwards in the Late Permian to Early Triassic (Fergusson 1991; Fielding *et al.* 1997).

Although igneous activity in both arches commenced in the Tournaisian (represented by *c.* 350 Ma granites), the main pulse of granite formation was in the Namurian stage, from *c.* 324 Ma until 313 Ma (predating uplift) in the Auburn Arch, and *c.* 316–305 Ma in the Connors Arch, where uplift was only minor (Murray 2003). Bryan *et al.* (2001) suggested that all these granites were extensional in character. Murray (2003), however, indicated that the older granites in the Auburn Arch and the older part of the Connors Arch are subduction-related, and that younger granites with large volumes of rhyolitic to dacitic ignimbrites and local andesite lavas in the Connors Arch spanned the changeover from subduction to the beginning of extension, around 305 Ma, the Carboniferous–Permian boundary (Murray 2003).

The Early Tournaisian–Viséan cycle of fill of the Yarrol Trough forearc basin (back-arc basin of Bryan *et al.* (2001, 2003*b*) was marked by the presence of felsic, rather than the intermediate, volcanic detritus of the first cycle, and by the presence of finer-grained sediments consisting mainly of fine-grained sandstone, siltstone and oolitic limestone (Yarrol Project Team 1997; Fordham *et al.* 1998). Granite clasts appear in the Yarrol Trough around the Viséan–Namurian boundary (*c.* 327 Ma) and reflect rapid exhumation of the arc to the west (Murray 2003).

The Yarrol Trough passes outboard across the Yarrol Fault System (Figs 12, 13) into ultramafic to mafic rocks of the Marlborough terrane in the north and into subduction complex rocks in the south (Fig. 13a). Subduction complex rocks in the northern New England Orogen comprise the North and South D'Aguilar and Beenleigh blocks in the south and the Wandilla and Shoalwater terranes in the north (Fig. 13a). The Beenleigh Block consists of greywacke, argillite and Early Carboniferous chert (Aitchison 1988), which are organized into packets of coherent turbidites separated by zones of broken formation (Smith 1999). Similar strata occur in the South D'Aguilar Block: both are interpreted as forming in upper levels of an accretionary complex (Holcombe *et al.* 1997*b*). The North D'Aguilar Block, in contrast, is a composite terrane, containing high-level accretionary complex rocks, as well as ophiolitic components of an accretionary complex that were subducted to more than 18 km before 315 Ma, and were then exhumed in the lower plate of a latest Carboniferous, low-angle normal fault (Little *et al.* 1992; 1995; Holcombe *et al.* 1997*a*). Further north, the Early Carboniferous Wandilla terrane consists mainly of bedded and melanged deep-water sandstone and mudstone. The volcaniclastic character of sandstone beds (with clasts of basalt, andesite and rhyolite) together with rare ash-fall tuffs suggests a provenance from the coeval magmatic arc and its basement to the west (Leitch *et al.* 2003). In contrast, the outboard Shoalwater terrane of presumed late Early Carboniferous age is dominated by craton-derived quartz-rich sandstone inferred to have been derived by longitudinal transport from the North Queensland Orogen (terminology of this paper) (Leitch *et al.* 2003).

In the southern New England Orogen, a *c.* 400 km belt of NW-trending Carboniferous volcanic rocks represent the outboard parts of the continental margin Currabubula arc that was developed on crust of the Lachlan Orogen (McPhie 1987) (Figs 12c, 13a). Volcanism reached a peak in the Viséan (342–327 Ma), after a minor pulse of volcanism in the Tournaisian (Fig. 4b) and was largely rhyolitic in the southern *c.* E–W-trending part of the arc, but mixed andesitic–rhyolitic further north. In both areas, volcanism continued locally into the early Stephanian (*c.* 303 Ma), but was rhyolitic–dacitic in both areas (Roberts 1995; Roberts *et al.* 2003; 2004; J. Roberts unpublished data) (Fig. 4b). The main volcanic products are low-K calc-alkaline rhyolitic ignimbrites with lesser calc-alkaline andesites (McPhie 1987; Roberts *et al.* 2003) that show LREE enrichment, moderately negative Eu anomalies, HFSE depletion and LILE enrichment (Jenkins *et al.* 2002). These features are all indicative of subduction zone magmatism. Highly radiogenic Sr suggests incorporation of continental crust into the melts (Jenkins *et al.* 2002).

These Carboniferous arc rocks are similar in age and chemistry to *c.* 340–320 Ma I-type granites in the Eastern subprovince of the Lachlan Orogen that are post-collisional to the Kanimblan cycle (Figs 6, 12a). Shaw & Flood (1993) suggested that the granites were deeper levels of a single arc that had been displaced and exhumed, whereas Jenkins *et al.* (2002) suggested that the granites represented a second arc. An alternative view is that the two rock packages are unrelated and the arc-like signature of the Carboniferous granites reflects partial melting of the underlying Ordovician arc volcanic rocks (Watkins 1998*a*; Glen 1998).

The Carboniferous stage of the forearc basin development is marked by major segmentation of the Tamworth Trough by cross-faults, represented in some cases by submarine channel margins (Roberts & Engel 1987; Roberts & Glen unpublished). Granite clasts in the Early Viséan (*c.* 340 Ma) and the base of Namurian (*c.* 327 Ma) were sourced from the exhumed arc to the west (J. Roberts pers. comm. 2004). Deposition ceased at *c.* 305 Ma (Roberts & Geeve 1999).

Most of the accretionary complex in the New England Orogen is Carboniferous in age (Figs 4b, 4c), with ages originally based on linking oolitic detritus back to Early Carboniferous oolitic limestones in the forearc basin (Murray 1997). This complex was called the Texas Terrane by Cawood & Leitch (1985) and corresponds largely to the Anaiwan Terrane of Aitchison *et al.* (1992*a*). High-level, frontal accretion is recorded by imbrication of turbidite sequences and formation of melanges of Early and Late Carboniferous age (Fergusson 1985). Aitchison *et al.* (1992*a*) showed that the typical stratigraphy of a fault slice consisted of basal mafic volcanic rocks, Late Devonian cherts, Tournaisian tuffaceous siltstone and chert interbedded with sandstone passing up into the dominant Tournaisian–Viséan volcanogenic sandstone (Fig. 11b). The presence of these volcanogenic sandstones indicates that the younger Carboniferous part of the accretionary prism developed relatively close to the continental margin arc (Aitchison *et al.* 1992*a*).

Subduction underplating is recorded by the presence of riebeckite-bearing terranes, now exhumed along major faults (Dirks *et al.* 1992). K–Ar riebeckite ages of 319 Ma and 312 Ma reflect high-pressure subduction metamorphism (Watanabe *et al.* 1988).

Deformation. In the southern New England Orogen, the cessation of subduction was followed by a major *c.* 311–300 Ma event that culminated in uplift and emplacement of early granites at 302 Ma and 300 Ma (Hillgrove suite and Wongwibinda complex, respectively: Dirks *et al.* 1992; Landenberger *et al.* 1995). The early fabrics in these metamorphic complexes have been interpreted as contractional (Collins *et al.* 1993). Approximately synchronous cessation of deposition in the forearc basin at *c.* 305 Ma (Roberts & Geeve 1999) was followed by major strike-slip faulting and anticlockwise rotation of crustal blocks. Three blocks in the southern part of the Tamworth Trough underwent Late Carboniferous, palaeomagnetically-determined, anticlockwise rotation (Geeve *et al.* 2002), although the contribution of thrusts to vertical-axis rotation has not yet been determined. A fourth block, the Hastings Block, underwent *c.* 250 km of left-lateral translation to the NE as well as anticlockwise rotation (Scheibner 1976; Cawood 1982; Cawood & Leitch 1985; Schmidt *et al.* 1994; Collins *et al.* 1993; Roberts & Geeve 1999) (Figs 12c, 13a). These rotated blocks were then folded, cleaved and faulted in the Late Carboniferous, before deposition of disconformably-overlying Early Permian strata and intrusion of the Barrington Tops Granite (which has a U–Pb zircon date of 281 Ma, Kimbrough *et al.* 1993).

Carboniferous and Devonian rocks in the northern part of the Tamworth Trough were telescoped by folding, variable cleavage development and thin-skinned thrusting into a west-vergent foreland fold thrust belt (the Tamworth Belt) that defines the external part of the New England Orogen (Glen & Brown 1995; Woodward 1995). This deformation was probably also Late Carboniferous, since Allan & Leitch (1990) recorded 4000 m of pre-Permian uplift west of the Peel Fault System and pre-Permian low-grade regional metamorphism, deformation and erosion to the east. The similarity of detritus in Early Permian stratigraphic units across the Peel–Manning Fault System suggests that the forearc basin and subduction complex had been largely juxtaposed by the Early Permian (Allan & Leitch 1990; Dirks *et al.* 1992).

In Queensland, uplift in the arc is Late Carboniferous, around 305 Ma (Hutton *et al.* 1999), and postdates mid-Late Carboniferous inversion of the back-arc Drummond Basin that produced folds and thick-skinned oblique thrusts with N–S trends in the south but which wrap around the Anakie Inlier in the north (Johnson & Henderson 1991). Deformed rocks are overlain unconformably by Late Carboniferous volcanic rocks and granites (Caritat & Braun 1992).The cause of this Late Carboniferous deformation is uncertain: it could reflect 'hiccups' in steady-state subduction, regional deformation (e.g. Allan & Leitch 1990), or it could be extension-related, such as inferred in the northern New England Orogen.

Cycle 3: Early Permian

Extension. In the New England Orogen, latest Carboniferous to Early Permian extension is recorded by (1) emplacement of granites and serpentinites into, and formation of low-angle extensional faults in, the former accretionary prism; and (2) formation of rift sedimentary basins.

(1) Granites were intruded into Devonian and Carboniferous accreted terranes in the Late Carboniferous (Fig. 13a). In the North D'Aguillar Block, in the northern New England Orogen, these granites were emplaced at *c.* 306 Ma during regional extension that led to generation of a major low-angle normal extensional fault. This fault separates multiply-deformed, blueschist–greenschist grade rocks in the lower plate from prehnite-pumpellyite grade, high-level accreted rocks in the upper plate (Little *et al.* 1992; 1995). In the southern New England Orogen, the Tia Complex was uplifted at *c.* 311 Ma (Dirks *et al.* 1992) and intruded by granite at 302 Ma. Uplift was thought to be contractional (Collins *et al.* 1993) but could be extensional in nature. The emplacement of serpentinite along the Peel–Manning Fault System in the Early Permian (Leitch 1969), during regional extension is recorded by the presence of clasts of serpentinite and sedimentary serpentinites in the Permian (Cross *et al.* 1987; Allan & Leitch 1990). It is also consistent with Early Permian (286 and 279 Ma) K–Ar ages of nephrite from a splay fault of the Peel Fault System (Lanphere & Hockley 1976).

(2) Early Permian rift basins are widespread (Fig. 14a), and were developed above deformed rocks of the forearc basins, the accretionary complexes, as well as on top of rocks of older basement (the Lachlan, Thomson and North Queensland orogens) to the west. Early Permian basins above accretionary terranes in the northern New England Orogen have been described by Sliwa *et al.* (1993). In the southern New England Orogen, these basins are either structural remnants of one large basin (Barnard Basin of Leitch 1988) or formed as separate narrow fault-bounded transtensional basins along major faults such as the Peel–Manning Fault System (Korsch 1982; Flood & Aitchison 1992). Vickers (1999) gave one example of a local basin formed during dextral overall transpression. The Gloucester Basin in the southern New England Orogen developed above deformed rocks of the Tamworth Belt. Basal units consist of basalts and andesites with some rhyolites (the *c.* 275 Ma, Alum Mt Volcanics) that have MORB-like affinities with some input either from a crustal or subduction modified source (Roberts *et al.* 1996).

Extension of crust of the North Queensland Orogen is reflected by generation of small Early Permian rift basins (Davis & Henderson 1999) and emplacement of widespread felsic volcanic rocks and granites (Bain & Draper 1997). Extension of Lachlan and Thomson crust west of the New England Orogen is represented by the Early Permian inception and growth of the Sydney, Gunnedah and Bowen basins as extensional, or more likely, transtensional rifts that are floored by volcanics or underlain by intrusive rocks such as represented by the regional elongate Meandarra gravity ridge (Qureshi 1984; Murray *et al.* 1989). A combination of regional sag (post-rift subsidence) and cessation of loading in the late Early Permian led to establishment of marine conditions that prevailed in more distal parts of the basin system. Early deformation in the western Bowen Basin produced a gentle inversion unconformity (Korsch *et al.* 1998).

Rift volcanic rocks at the base of the Sydney Basin include the 292 Ma (Rb–Sr) Rylstone

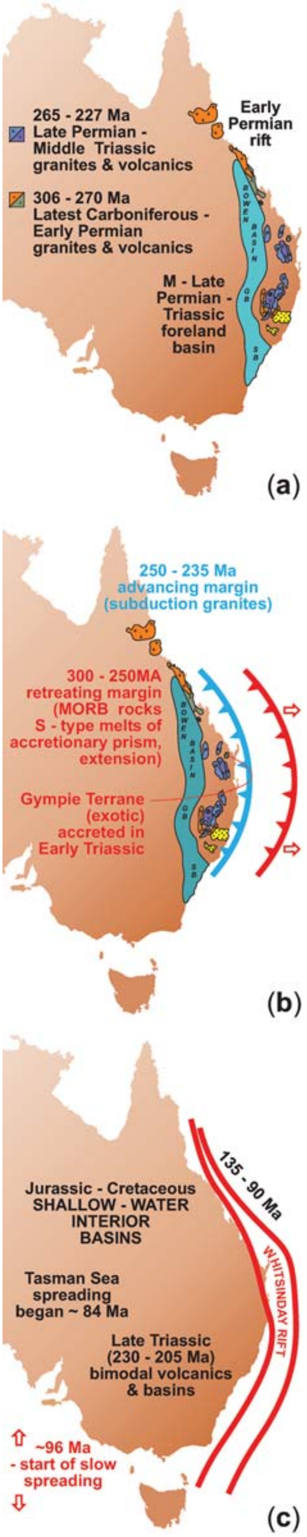


Fig. 14. (**a–b**) Hunter–Bowen supercycle: (**a**) cycle 3 showing igneous activity and Bowen, Gunnedah and Sydney basins that developed from rift (transtensional) basins into foreland basins; (**b**) Permian to Triassic advancing and retreating margins (after Jenkins *et al.* 2002); (**c**) Mesozoic rifting and basin formation.

Volcanics near the western margin (Shaw *et al.* 1989) and the 270 Ma Werrie Basalt (with remnant magnetism) that lies at the base of the NE Sydney Basin as well as underlying most of the Gunnedah Basin (Leitch 1993; Caprarelli & Leitch 2001). This basalt has a mixed OIB and subduction-related signature, possibly reflecting derivation from melting of mantle enriched during Carboniferous subduction (Caprarelli & Leitch 2001). High level *c.* 270 Ma igneous rocks intrusive into the Werrie Basalt have subalkaline to alkaline chemistry. They were derived from depleted mantle with some crustal contamination and carry 340 Ma and 300 Ma detrital zircons (Teale *et al.* 1999).

Silicic flows and ignimbrites of the Boggabri Volcanics (Leitch 1993) outcrop as a curved NW belt near the eastern margin of the Gunnedah Basin and extend westwards at depth under most of the basin (Leitch 1993; Tadros 1993; Beckett *et al.* 1995). They have a predominant MORB-like geochemistry, but with a subduction signature inherited from Carboniferous subduction during Early Permian extension (Brownlow 1999; Brownlow & Arculus 1999).

Volcanic rocks at the base of the Bowen Basin in Queensland include the Rookwood Volcanics, bimodal rocks of the Early Permian Lizzie Creek Volcanics and the dominantly intermediate and felsic rocks of the Late Carboniferous (Stephanian) to early Permian Camboon Volcanics (Hutton *et al.* 1999; Fielding *et al.* 2001).

Interpretations of the Early Permian extension in the Hunter–Bowen supercycle include subduction of a spreading ridge (Murray *et al.* 1987), slab break-off (Caprarelli & Leitch 1998), and changes in the Gondwana–proto-Pacific plate boundary configuration from retreating to advancing (Jenkins *et al.* (2002) (Fig. 14b).

Orocline formation. The southern New England Orogen is dominated structurally by three orogen-scale folds or oroclines in accretionary complex rocks; the southern one is also outlined by the curve in the Peel–Manning Fault System (e.g. Korsch & Harrington 1987; Murray 1997)

(Figs 12c, 13). The southern (Manning) orocline began to form in the Late Carboniferous with rotation and left-lateral translation of blocks in the Tamworth Trough described above, and formation of radiating faults (e.g. Korsch & Harrington 1987). It continued into the Permian, since Permian basins were caught up in a N–S deformation at *c.* 268–267 Ma (Roberts & Geeve 1999). This sense of faulting is consistent with left-lateral movement on the Peel–Manning Fault System to the northwest, where it is recorded by oblique S2 cleavage in the eastern part of the Tamworth Belt, transecting early D1 folds (e.g. Cao & Durney 1993; Glen unpublished data), and by strike-slip fabrics in schistose serpentinite east of the Peel Fault System east of Manilla (Glen & Brown 1995). The northern (Texas) orocline also deformed Early Permian sediments (Lennox & Flood 1997) and Korsch & Harrington (1987) showed that this involved right-lateral strike-slip on the eastern limb. Together the northern and southern oroclines define a long doubly-plunging regional (500 km) antiform. The eastern (or Coffs Harbour) orocline formed when accretionary complex rocks of the Anaiwan/Texas Terrane to the north were translated southwards, against the northwards-moving Hastings Block, also resulting in east–west structures in the Early Permian Nambucca Block (Johnston *et al.* 2002). Subsequent east–west shortening of these rocks at 266–258 Ma is dated by uplift on bounding faults to the west (Dirks *et al.* 1992; Landenberger *et al.* 1995).

Formation of these oroclines by synchronous yet oppositely directed strike-slip faulting seems to be a paradox. A possible explanation is suggested here, involving major out-of-plane movement, with the almost complete closure of accretionary complex rocks representing a section through a steeply plunging non-cylindrical antiform. In this model, the core of the fold underwent sub-vertical extension, bringing high-grade accretionary complex rocks to the surface. This uplift corresponds to the 310–300 Ma S3/L3 event of Dirks *et al.* (1992). The large scale of this structure (*c.* 500 km) is consistent with uplift of 10–15 km (Dirks *et al.* 1992). As part of this extension, outer, low-grade rocks of the Tamworth Belt underwent opposite-directed translation and rotation, with radiating faults focused on the southern orocline. Such a 'sheath fold' model invites comparison with the core complexes described from Queensland by Little *et al.* (1992).

Cycle 4: Late Early Permian–Triassic

Convergence. Renewed convergence in the late Early Permian to Triassic was accompanied by the extrusion of widespread ignimbrite sheets and emplacement of major I-type Late Permian–Late Triassic (255–222 Ma) granites of the New England Batholith. Most of these granites lie in the southern New England Orogen (Figs 13a, 14). Shaw *et al.* (1991) recognized three groups of granites and associated volcanic rocks, with most emplaced between 253 Ma and 244 Ma, some in NNE-trending graben or rifts. Some granites extend into the northern New England Orogen, where they range in age between 270 Ma and 205 Ma (Murray 2003). These granites are high-level and mainly I-type, although some A-types are present, as are zoned granites and layered gabbros (Murray 2003).

Cycle 4 granites have a close geochemical similarity with Mesozoic granites of the North American Cordillera (Cawood 1984; Chappell 1994; Bryant *et al.* 1997), indicating that they are subduction-related, especially those of the Clarence Supersuite in the east (Bryant *et al.* 1997), which extend into the northern New England Orogen. Gust *et al.* (1993) also argued for a subduction-related origin, but suggested that there was a transition to extension in the Late Triassic due to slab rollback. Chappell (1994) noted that the I-type Moonbi suite in the southern New England Orogen was K-rich, like the Kanimblan post-collisional Bathurst granites in the eastern Lachlan Orogen. Could they also be melts of Ordovician volcanic rocks in the under-thrust Lachlan Orogen (see below)?

The Gympie Terrane (or Gympie Province, Cranfield *et al.* 1997) (Figs 13a, 14b) in the outboard part of the northern New England Orogen is a key component of cycle 4 (Harrington 1974). This terrane is in fault contact with the Carboniferous accretionary prism to the west and consists of Early–mid-Permian volcanic rocks passing up into Late Permian limestone and then into Early Triassic continental, then marine, units (Cranfield *et al.* 1997; Sivell & McCulloch 2001). This terrane is interpreted as a primitive oceanic arc associated with enriched subcontinental lithosphere, subsequently rifted in the Early Permian. Proximity to the Gondwana margin is debatable (Harrington 1983; Cranfield *et al.* 1997; Sivell & McCulloch 2001; Holcombe *et al.* 1997*b*). There are similarities with the offshore South Island terrane (Fig. 13a), which consists of mafic and ultramafic rocks with a maximum age of *c.* 277 Ma (Early Permian) and which formed in an intra-oceanic arc (Bruce & Niu 2000*a*). Similar-age volcanic elements in the Berserker terrane (erupted in a back-arc or intra-arc setting, Crouch 1999) and in the South Island terrane suggest relative proximity (Bruce & Niu 2000*a*).

West of the New England Orogen, cycle 4 was reflected by conversion of rift basins in the Bowen–Gunnedah–Sydney basin system to foreland basins (Figs 13a, 14). Foreland loading by the southern New England Orogen first commenced in the late Early Permian and is recorded by deposition of the oldest coal measures (Greta Coal Measures) mixed with volcanic detritus in the NE part of the Sydney Basin. Syn-sedimentary thrusting in the NE Sydney Basin led to the formation of growth anticlines and westward-propagating thrust fronts (Glen & Beckett 1997) and produced southward-draining axial palaeodrainage ahead of the thrust front. In the Gunnedah Basin, equivalent deposits in the eastern Maules Creek sub-basin were derived from basal volcanic rocks upthrust to the west (Beckett *et al.* 1995). Minor Late Permian (*c.* 252 Ma) anticlinal growth further west is reflected by flat-lying earliest Triassic rocks above dipping Permian reflectors (Tadros 1993). In the Bowen Basin, thrusting produced the Gogango Overfolded Zone (Fergusson 1991; Fielding *et al.* 2001).

By the end of the Late Permian (Tartarian), all three basins had been converted to coal-bearing foreland basins, fed in pulses from first cycle volcanic detritus and uplifted detritus (e.g. jaspers from the accretionary complex) from the New England Orogen in the north and east (Fielding *et al.* 2001), where crustal loading was synchronous with active volcanism and granite formation. Tuffs in Late Permian coal measures reflect coeval volcanism in the southern New England Orogen (Brownlow 1979; Shaw *et al.* 1991), with the oldest tuffs dated at *c.* 264 and *c.* 265 Ma. Mixed volcanic and quartz-rich detritus in some stratigraphic units, especially in the Gunnedah Basin, reflects input from both the New England Orogen and a westward-receding crustal bulge in the older Lachlan Orogen to the west.

A major environmental change just above the Permian–Triassic boundary (*c.* 250 Ma) coincided with the end of tuff and coal deposition and the onset of red-bed oxidized fluviatile sedimentation of the Rewan Group in the Bowen Basin and the Narrabeen Group in the Sydney Basin (Veevervood). The last unit especially was derived from the New England Orogen and was synchronous with major periods of volcanism and granite emplacement. The thrust front reached the eastern margin of the Bowen Basin during this event.

This Early Triassic phase of basin filling lasted only *c.* 14 million years (250–244 Ma) and was followed by relative uplift in the craton from 244 Ma to 235 Ma (Fielding *et al.* 2001), leading to the shedding of large amounts of craton-derived detritus into the Bowen, Gunnedah and Sydney basins to form the Clematis and lower Napperby formations and the Hawkesbury Sandstone, respectively. Final filling of these basins (235–230 Ma) was dominated again by detritus shed from the New England Orogen (which included coeval tephra deposits, Shaw *et al.* 1991), although the finer-grained nature of sediments, commonly lacustrine, point to lesser topographic expression and reduced thrusting.

Collision. In the New England Orogen, cycle 4 was terminated by the collision of the Gympie Terrane/Province with Gondwana in the Early to Middle Triassic (Sivell & Arnold 1999) or Middle to Late Triassic (Cranfield *et al.* 1997). This deformation is inferred to have caused the Bowen phase of the Hunter–Bowen Orogeny, a major crustal loading event in the New England Orogen (Cranfield *et al.* 1997). The Sydney, Gunnedah and Bowen basins became converted into fold–thrust belts as the deformation fronts migrated westwards. An Early Triassic contractional deformation is recorded from the Bowen Basin and a Middle Triassic event from the Drummond Basin to the west (Johnson & Henderson 1991), before the terminal Late Triassic deformation at *c.* 233 Ma (Korsch *et al.* 1998), which also involved rocks of the Tamworth Belt being thrust westward over Middle Triassic and unconformably-overlying Early Jurassic rocks of the Surat Basin (Wartenberg *et al.* 1999).

In central Queensland, widespread thrusting and open folding in the central and western parts of the Bowen Basin suggest greater amounts of crustal shortening than in the Gunnedah and Sydney basins. Indeed, Malone (1964) and Fielding *et al.* (1997) suggested that the basin extended up to 200 km to the east, reaching the Queensland coast before deformation, with only small relics preserved on the surface and in the subsurface. In contrast, the Gunnedah Basin underwent less shortening below the bounding 15° E-dipping Mooki Thrust, so that only the eastern parts (the Maules Creek Sub-basin and Boggabri Ridge) in the footwall underwent significant thrust-related deformation (Beckett *et al.* 1995). The western part of the Gunnedah Basin still retains elements of its early rift geometry, being divided into north–south blocks by major cross-faults (Tadros 1993). Further south, the northeastern margin of the Sydney Basin lies in the lower plate of the major east-dipping Hunter Thrust, which has a complex history of greater shortening in the north (Glen & Beckett 1997; Glen & Roberts unpublished data). Thrusting in the offshore part of the New England Orogen was west-directed and is probably responsible for formation of major N–S folds and blind faults in the main part of the Sydney Basin.

Hunter–Bowen cycle 4 also affected basement west of the New England Orogen, with the emplacement of large amounts of granite and felsic volcanic rocks in the North Queensland Orogen (Bain & Draper 1997). In the North Queensland Orogen, the Hodgkinson subprovince that underwent D3 vertical shortening during cycle 3 extension (Davis & Henderson 1999) underwent D4 horizontal crustal shortening in cycle 4, which is best constrained by the syn-deformational intrusion of S-type granites (dated at 280–265 Ma, Davis *et al.* 2002). Subsequent formation of small Early–Late Permian coal basins were deformed in the Late Permian–Early Triassic by the Bowen phase of the Hunter–Bowen Orogeny (Bain & Draper 1997).

Mesozoic rifting

After the end of the Hunter–Bowen supercycle, the plate margin was relocated to the east. As a consequence of the opening of the Tasman Sea, beginning at *c.* 90 Ma, the record of this younger Mesozoic interaction between Gondwana and the proto-Pacific plate is preserved in the islands of the southwest Pacific and in New Zealand. The Australian sector of Gondwana underwent Late Triassic rift-related volcanism and slow subsidence that led to the formation of widespread Jurassic–Cretaceous shallow-water basins. The prominent Jurassic Whitsunday Rift developed along the eastern margin of the continent from *c.* 135 Ma to 90 Ma (Bryan *et al.* 1997) (Fig. 14c).

Part 3: Discussion

In this third part of the paper, some of the more keenly-debated parts of the history of Delamerian and Lachlan cycles are discussed from the point of view of both the author as well as other models.

Significance of cycles

Recognition of cycles in the Tasmanides emphasizes the wide-scale nature of stages, both extensional/magmatic/sedimentary and contractional/strike-slip, in the evolution of the Australian sector of East Gondwana. By drawing attention to linkages between orogenic belts, the recognition of cycles forces a wider-scale look at the growth of the Tasmanides. Without being too doctrinaire in this approach, recognizing the possibility of diachronous events and lack of correlations, several points can be made. Recognition of elements of the Delamerian cycle in the New England Orogen leads to questions on what happened in the outboard part of the Tasmanides during Rodinia break-up, and when subduction began (e.g. Cawood & Leitch 1985; Cawood 2002; Crawford *et al.* 2003*a*). The recognition of Lachlan cycles in the Delamerian Orogen suggests that cratonized crust and upper mantle can still respond to deformational and heating events generated at the plate margin further outboard. The application of Lachlan cycles to the North Queensland Orogen implies broad similarities in plate events, despite the fact that, as discussed below, one area underwent major back-arc spreading not evident in the other. If these two orogens were part of the same, albeit segmented, plate margin, where is the intervening part of this convergent margin? Was it overthrust by the outboard (Delamerian) part of the Thomson Orogen? Although Devonian elements of the Hunter–Bowen supercycle are interpreted as having affected the Thomson and North Queensland orogens, it is not clear how to separate them from elements of the Tabberabberan cycle. Finally, the presence of Delamerian to Tabberabberan cycles in the New England Orogen bears on the question of how allochthonous that orogen was in relation to the Gondwana margin. It suggests formation not too far from the Lachlan Orogen in the Ordovician–Late Devonian, an issue addressed most recently from palaeomagnetic data by Klootwijk (2002).

Rifting in the Delamerian cycle

The rift history of the Delamerian cycle represents the response to the break-up of Rodinia and the separation of Laurentia (e.g. Dalziel *et al.* 1994; Powell *et al.* 1994) although Direen & Crawford (2003*a*) preferred a model in which microcontinental ribbons were calved off. Most of our knowledge of the Delamerian cycle comes from the inboard parts of the

Tasmanides. Several issues can be addressed from these areas, such as the timing of rift–drift transition, the geometry of rifting and the relationship between Tasmania and the mainland before *c.* 700 Ma.

However, clues as to what happened further outboard during the *c.* 200 millions of years of rifting are less easy to find. Neoproterozoic rift volcanic rocks in the Anakie Inlier lie thousands of kilometres east of the Gondwana margin, as now accepted, and indicate that either our ideas of the shape of the East Gondwana margin were wrong or that its current location reflects subsequent rifting to the west. Similarly, the Marlborough Ophiolite in the New England Orogen indicates that seafloor spreading was underway in the Neoproterozoic. This is consistent with the presence of old lithosphere under the southern New England Orogen. This lithosphere could have rifted away from cratonic Australia, to be subducted early in Tasmanides history, or was brought in during subsequent Palaeozoic to Mesozoic subduction or strike-slip faulting. Part of the answer to this question may come from western Tasmania, which shows an early history not recorded on the mainland (e.g. significant rift tholeiites >780 Ma) and which appears to have lain further outboard than the Adelaide Rift Complex before deposition of *c.* 700 Ma Sturtian tillites (e.g. Elliott & Gray 1992; Powell *et al.* 1994). Berry *et al.* (2001) showed that the detrital zircon pattern in Tasmanian Cambrian rocks differed from those of the Kanmantoo Group on the mainland and suggested that western Tasmania rifted from Australia and remained as a separate terrane or promontory until the Delamerian Orogeny.

Returning to the inboard part of Gondwana, the old edge of Rodinia can be reconstructed partially as a steep east-dipping normal fault (the Torrens Hinge Zone) on the western side of the Adelaide Rift Complex in the southern part of the Tasmanides, because of the limited reactivation of this zone by subsequent thrusting. Rifting occurred over 300 million years (827–527 Ma), and was taken up mainly by rift cycle 1 of this paper. It is generally agreed that the Australian craton was part of the Rodinia supercontinent and that rifting marked the opening of the proto-Pacific Ocean and separation of Laurentia (as well as the South China Block, Li *et al.* 1995) from Gondwana. Powell *et al.* (1993*b*) used palaeomagnetic data to suggest that this separation began after 725 Ma; newer data (Wingate & Giddings 2000) indicated that rifting occurred earlier, by *c.* 755 Ma. If so, this event was not only amagmatic, it was not reflected in the Adelaide Rift Complex for another 50 million years. The Tapley Hill Shale (post-Sturtian tillite) is the first stratigraphic unit that was deposited right across the rift system and interpreted by Powell *et al.* (1994) as reflecting the rift–drift transition. If separation occurred by 755 Ma the rift margin must have lain further east. What happened before 600–580 Ma, a period of major igneous rifting with production of continental rift alkaline–tholeiitic basalt and picritic volcanism, documented by Crawford and co-workers (e.g. Direen & Crawford 2003*a*)? Was the 600–580 Ma event linked to subsequent rifting of smaller continental blocks in newly formed Gondwana, calving off ribbons, or was it the main rift event, as argued by Direen & Crawford (2003*a*) and by Veevers *et al.* (1997), who suggested that the break-up of Rodinia occurred around 560 Ma.

The direction of rifting in event 1 is provided by the orientation of the Gairdner Dyke Swarm and the palaeogeographical reconstructions of the Adelaide Rift Complex by Preiss (2000). Both imply a NE direction of maximum extension from *c.* 827 Ma, in cycle 1 events 1–2, 4 (see the section on 'Delamerian cycle: extension and passive margin phase') until the Cambrian. Mixed NE and E extension characterized events 3 and 5, with east–west extension especially common in the southern part of the Adelaide Rift Complex, with the Torrens Hinge Zone initiated in cycle 1c as the western edge of rifting (Fig. 3b).

Significance of the Tasman Line

The Tasman Line was originally defined by Hill (1951) as the line marking the eastern margin of outcropping Precambrian rocks in Australia. Harrington (1974) extended the line to the south, and locations of these lines and other authors' versions are shown in Figure 1. Subsequently, the concept of the Tasman Line has been broadened to represent the western boundary of Tasmanides, with the line considered to mark the place of break-up of a Mesoproterozoic supercontinent of which Australia was part (e.g. Veevers & Powell 1984; Powell *et al.* 1994; Scheibner & Basden 1998; Scheibner & Veevers 2000; but not Direen & Crawford 2003*b*) (Fig. 1).

While the Tasman Line coincides with the western margin of the Tasmanides in North Queensland, it is a young contractional fault that adds little hard data to the debate about supercontinent break-up geometry or kinematics. South of the Diamantina River Lineament, most authors run the Tasman Line east of, and around, the Curnamona craton, which includes the Palaeo-Mesoproterozoic Willyama Supergroup rocks of Olary and Broken Hill, and east of the Neoproterozoic Adelaide Rift Complex (Fig. 1). However, these differing definitions of the Tasman Line do not differentiate between Neoproterozoic fill of a rift system (and its thinned basement of older thinned Palaeo- and Mesoproterozoic rocks) and unthinned basement to the west. It has been argued above that the true boundary lies further west, along the Torrens Hinge Zone (Fig. 1) (also Mills 1992). The Tasman Line has also been recognized in Tasmania as the Tamar Fracture along the NNW-trending Tamar River, separating eastern and western Tasmania (Williams 1978). This fault has been linked to the mainland by a WNW tear fault in Bass Strait (e.g. Scheibner 1974; Veevers & Powell 1984). Because geophysical modelling suggested that the Tamar Fracture has no crustal significance (Leaman 1994), Reed *et al.* (2002) recognized a new boundary, the west Tamar Fault, 20 km to the west, as an Early–Middle Devonian (Tabberabberan) NE-dipping thrust system (Fig. 1).

As a result of these points, it is suggested here that south of the Diamantina River Lineament, the Tasman Line has little tectonic significance (see also Direen & Crawford 2003*b*). This is consistent with an analysis of recent earthquake SKIPPY data that found little correspondence between the Tasman Line and the eastern edge of Proterozoic Australia, except along the Palmerville Fault System (van der Hilst *et al.* 1998).

Several authors have used the zig-zag shape of the southern part of the Tasman Line that wraps east around the Curnamona craton (Fig. 1) to suggest that either NE-trending segments formed as extensional faults (Veevers & Powell 1984; Powell *et al.* 1994) or that NW-trending segments formed as extensional faults (Powell 1998; Gibson 1998). These interpretations assume that the zig-zag in the Tasman Line around the Curnamona craton east of Broken Hill is an original feature. However, this shape may reflect later deformation that progressively plastered the Delamerian Orogen onto the older craton (Scheibner & Basden 1998; Scheibner & Veevers 2000), in the same way as the Fleurieu arc south of Adelaide was compressed onto the SW corner of the Gawler craton (most recently, Marshak & Flöttman 1996). If correct, the original configuration of the southern part of the Tasman Line was approximately planar without zig-zags. In this case, the locations of the 600–580 Ma rift volcanic rocks of Direen & Crawford (2003*a*) would restore to an approximately linear trend.

Convergence, collision and post-collision in the Delamerian cycle

The Delamerian/Tyennan Orogeny in western Tasmania and western Victoria was triggered by arc–continent collision around 510–505 Ma (Berry & Crawford 1988; Crawford 2003*a*), followed by deformation of 500 Ma post-collisional volcanic rocks and rift basins around 495 Ma. Outboard parts of the Delamerian Orogen in western Tasmania and western Victoria preserve evidence of accreted Cambrian forearc boninitic crust – the Tasmanian mafic–ultramafic complex (Crawford & Berry 1992) and the lower part of the Stavely Volcanic Complex, respectively. These underwent post-collisional extension leading to emplacement of andesitic rocks (Crawford 2003*a*). Using West Pacific analogues, Crawford (2003*a*) argued that rocks of the same association in hanging walls of major thrusts in the Lachlan Orogen represent forearc crust that was distal to the collision. The actual colliding arc originally lay further east, according to the Crawford model, although several authors (Scheibner 1989; VandenBerg *et al.* 2000; Cayley *et al.* 2002) have suggested this accreted arc lies in western Victoria.

Calc-alkaline volcanic rocks in the hanging wall of the Mt Wellington Fault are more ambiguous. They could represent part of the (largely subducted) colliding arc, but they are probably post-collisional, like the Mt Read Volcanics in western Tasmania that they resemble geochemically (Crawford *et al.* 1996). However, that means there must have been a Delamerian/Tyennan deformation that extended into Victoria, despite the lack of recognition of an earlier 'ophiolite' obduction event (VandenBerg *et al.* 1995; 2000; Crawford 2003*b*). VandenBerg *et al.* (1995, 2000) argued that these calc-alkaline rocks are windows into underlying west Tasmania crust – the Selywn Block of Cayley *et al.* (2002). There are two other possible interpretations. The least likely accreted is that these are calc-alkaline rocks of another arc. This is effectively precluded by the similarity in geochemistry with the Mt Read Volcanics and the Stavely Volcanic Complex (Crawford *et al.* 1996; A. Crawford pers. comm. 2004). The preferred view here is that while the andesites are geochemically post-collisional, using the arguments of Crawford, their position so far east of the Delamerian margin reflects post-Delamerian rifting. It is thus envisaged that pieces of a forearc crust and post-collisional volcanic rocks, accreted to the East Gondwana Orogen during and just after the terminal Cambrian deformation, were rifted away during Early Ordovician rollback of the plate boundary. Using the terrane model of Glen & Percival (2003) and Glen (2004), these calc-alkaline Cambrian volcanic rocks form local substrate to the Bendigo Terrane, which accumulated as submarine fan systems off Antarctica and which was translated north to be accreted to the Australian part of East Gondwana in the Middle Devonian (see below). In this interpretation, there would be no continental crust (Selywn Block) under central Victoria, and no widespread Cambrian oceanic crust under the Southwestern subprovince: only locally stranded rifted segments embedded in younger Ordovician oceanic crust.

The tectonic significance of the geophysically defined western volcanic belt, closer to the Gondwana margin, is more uncertain. It may represent an arc, possibly related to west-dipping subduction and accreted at *c.* 530 Ma, before formation of the Kanmantoo Trough wich underwent deformation beginning at 515 Ma and followed by post-collisional volcanism, just as inferred from the eastern volcanic belt by Crawford *et al.* (1996). Such a history is consistent with the early cleavage and foreland basin formation cited earlier. The 'rift or arc' interpretation of the *c.* 526 Ma Koonenberry andesites is an important part in reconstructing this story and, thus, the nature of the Delamerian margin.

The curvilinear, largely concealed, mafic and ultramafic rocks along the southern margin of the Thomson Orogen may indicate the presence of a third Delamerian arc, possibly mixed in with ocean crust igneous rocks from the inferred suture between the two orogens. If these rocks are Cambrian, then a major part of the convergent East Gondwana margin had a latitudinal trend and a north–south vector of subduction.

The Cambrian ultramafic and related volcaniclastic rocks along the Peel–Manning Fault System in northeastern New South Wales are also enigmatic, since their geochemistry

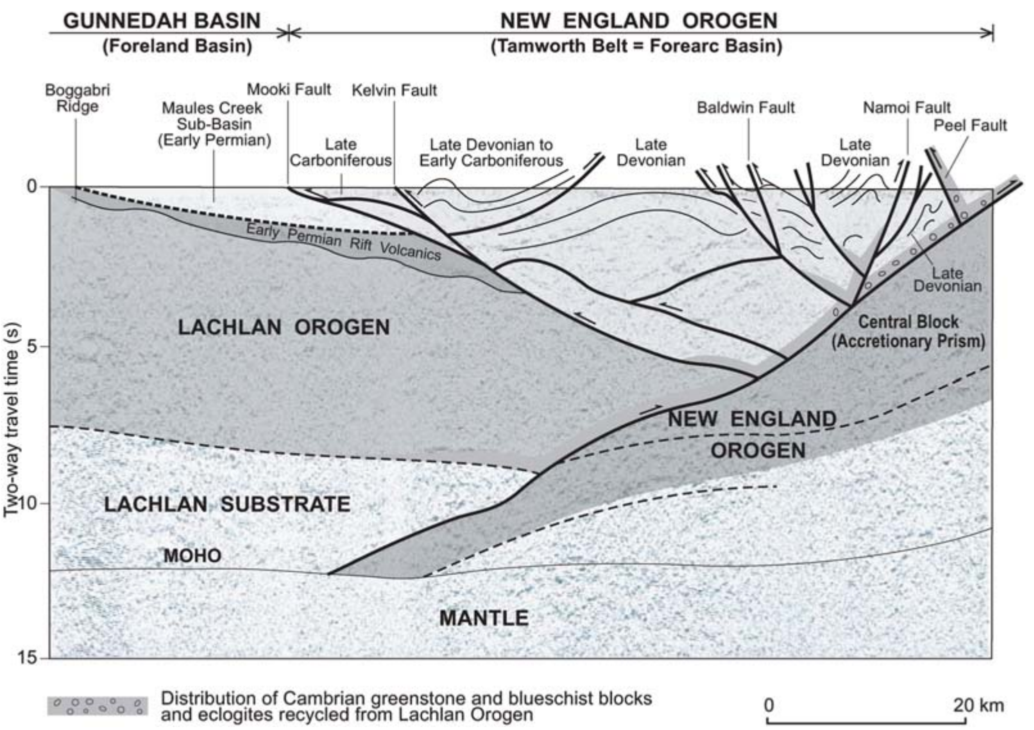


Fig. 15. Seismic section from Lachlan Orogen in the west into New England Orogen in the east (acquired by Geoscience Australia). Blueschists are found as knockers in schistose serpentinite along major faults such as the Peel Fault, part of the Peel–Manning Fault System. Based on Glen & Brown (1995) after Glen *et al.* (1993).

suggests strong affinities with forearc rocks in western Tasmania and Victoria. Their location along a major Permian fault system implies that they have been thrust up from beneath the forearc basin of the New England Orogen. It is suggested here that they are allochthonous and part of Cambrian igneous substrate to the Lachlan Orogen (Fig. 15) that was translated northwards as part of major plate boundary rearrangement around the Ordovician–Silurian boundary (see below).

Key features in the Lachlan supercycle

Benambran cycle

After the accretion of the old Delamerian Cambrian forearc crust, the Australian proto-Pacific plate boundary underwent rollback of *c.* 1000 km. This rollback is reflected by:

1. post-collisional extension, volcanism and uplift in the Delamerian Orogen;
2. splitting of the segments of the accreted Cambrian arc and forearc system that stretched from Antarctica to the New England Orogen (Münker & Crawford 2000). One rifted segment is inferred from geochemical data to constitute substrate to the Macquarie Arc (Glen *et al.* 2003; Crawford *et al.* 2005). It was suggested above that a second segment is represented by Cambrian calc-alkaline volcanic rocks of the Mt Wellington Fault zone in central Victoria. These volcanic rocks were covered progressively by clastic sediments, then deep-sea cherts, as they became isolated from the Gondwana margin;
3. formation of a wide back-arc basin (the Wagga Basin of Packham & Falvey 1971) in the Central and Western subprovinces filled by Early–Middle Ordovician turbidites. Earliest Ordovician turbidites are deposited on Cambrian igneous crust of the Delamerian cycle. The new Gondwana–proto-Pacific plate boundary was segmented (Fig. 16a).
   (a) The ‘strongly’ convergent northern segment, *c.* 1000 km long, formed opposite the Macquarie Arc. Behind the arc, the Wagga Basin developed as a back-arc basin floored by igneous crust and linked to the Macquarie Arc by the presence of Ordovician back-arc basin basalts. Further development of this basin led to deposition of Early–Middle Ordovician turbidites. These are overlain by a starved Late Ordovician black shale sequence (Colquhoun *et al.* 2004) that developed as the depocentre moved out of range of terrigenous sedimentation from the continental margin (Glen 2004). These black shales are interpreted as responses to renewed phase(s) of back-arc basin opening (Fig. 16b). New igneous substrate formed during this spreading is manifest as Ordovician MORB-like volcanic rocks – the Narragudgil Volcanics (Duggan 2000), the Nacka Nacka Complex (Basden 1990; Meffre & Glen unpublished data) and tholeiitic basaltic schists of the Tottenham Group with MORB chemistry (Muir 1999). These volcanic rocks now occupy high-strain zones or fault blocks with Early–Middle Ordovician turbidites. Narrow belts of schistose serpentinite are inferred to mark faults cutting down into this substrate.
   (b) The longer, southern segment was probably strike-slip or highly oblique (Glen & Percival 2003; Glen 2004; Glen *et al.* 2004) (Fig. 16a). It was marked by the presence of several large turbidite fan systems containing Early–Middle Ordovician craton-derived detritus. The Bega terrane was shunted to the north in the Late Ordovician to lie ‘outboard’ of the Macquarie Arc, and the Bendigo terrane was shunted northwards in the Silurian–Early Devonian (Fig. 16).

The Bega terrane now lies ‘outboard’ of the Macquarie Arc (Fig. 16f). Most of the terrane is occupied by Early–Middle Ordovician craton-derived quartz-rich turbidites that are faulted against volcanic rocks of the Macquarie Arc and show no sign of mixing of provenance (Glen *et al.* 1998; Meffre *et al.* 2005). While submarine systems can bypass arcs to deposit material on the incoming oceanic plate, as canvassed by Glen *et al.* (1998), this would still produce mixed provenance. Evidence from the Western Pacific suggests that this lack of provenance mixing constrains the two terranes to have formed at least hundreds of kilometres apart (Meffre *et al.* 2005). The Bega terrane is thus inferred to have been shunted northward along the plate boundary (Figs 16b–c), with the incoming of the condensed Late Ordovician black shale sequence reflecting the drift of the terrane away from the Antarctic margin towards the Lachlan Orogen (Fig. 16b–c) (Glen 2004). The Bendigo terrane remained anchored off Antarctica until the Silurian; alone of all the turbidite fan systems, it shows a west–east thinning of grain size and fining of rock packets consistent with increasing distance from the Gondwana margin.

The view that the Late Ordovician condensed black shales sequence in the Girilambone–Wagga terrane and Bega terrane represents periods of back-arc spreading and strike-slip along the Gondwana margin differs from that of Cas (1983), who suggested that the black shales represented deposition on submarine highs bypassed by turbidite distributary systems. Now that more is known about the areal extent of these shales, other origins must be sought – VandenBerg & Stewart (1992) suggested sea-level rise. However, Late Ordovician turbidites occur just east of Melbourne, so this explanation does not work. Fergusson & Fanning (2002) suggested that the black shales in the Bega terrane represent background sedimentation when terrigenous deposition was blocked by subduction of the Wagga Basin. However, black shales also represent the Late Ordovician fill of that basin, the Girilambone–Wagga terrane.

In this strike-slip model (Glen & Percival 2003; Glen 2004), Ordovician volcaniclastic rocks and blueschists along the Peel–Manning Fault System (and other faults) represent the missing forearc and accretionary prism rocks that originally formed outboard of the Lachlan Macquarie Arc (Fig. 16a). Cambrian ophiolites of the Delamerian cycle in the New England Orogen are similar chemically to those in the Lachlan Orogen (Aitchison & Ireland 1995) and were also caught up in this shunting, to be exhumed later along major faults (Fig. 16). This is somewhat similar to the idea of Cawood & Leitch (1985) that Ordovician volcaniclastic rocks along the Peel–Manning Fault System were part of the Ordovician Macquarie Arc, subsequently rifted off in the Silurian–Devonian. Such an interpretation is tenable, since the southern part of the New England Orogen lies offshore and outboard of the Lachlan Orogen (Fig. 2), but it is argued here that strike-slip is still needed to explain these Lachlan-type rocks in the northern part of the southern New England Orogen. In Part 1 of this paper, evidence was cited that the New England Orogen was underlain by old, fast lithosphere. Here, it is suggested that this lithosphere may correspond to the Cambrian–Ordovician parts of the Lachlan Orogen and its older substrate.

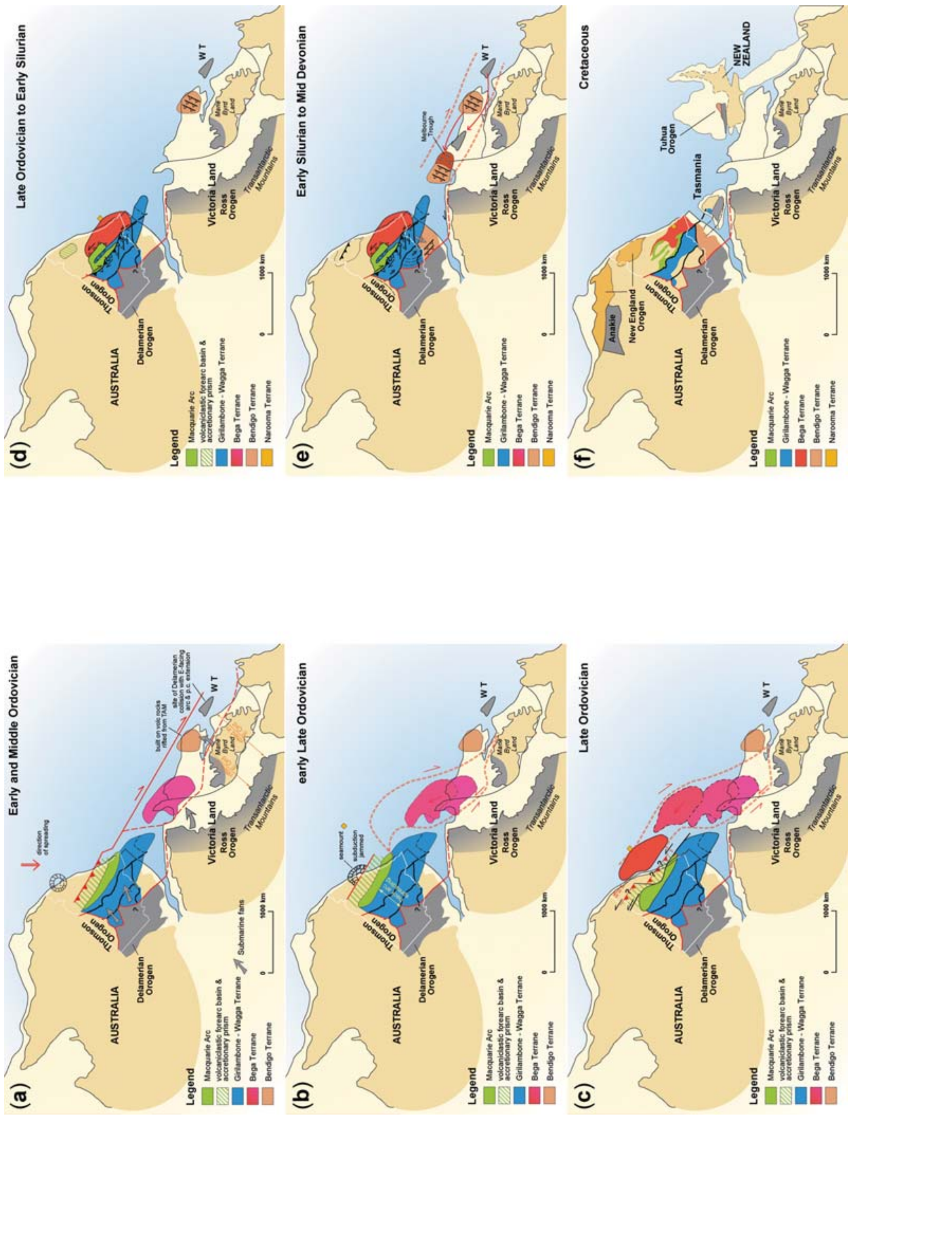
The two-stage Benambran Orogeny is explained by the Bega terrane entering the northern part of Gondwana–proto-Pacific plate boundary (Fig. 16c) and causing it to jam (Glen 2004). As part of this deformation, the Bega terrane was accreted to the Macquarie Arc, which was driven into the back-arc basin, the Girilambone–Wagga terrane (Fig. 16d). Most of the deformation was taken up in these turbidite terranes by formation of multiple cleavages, folds and oblique thrusts. Earliest Silurian sinistral strike-slip faulting inserted some back-arc Ordovician craton-derived turbidites into an 'inter-arc' position between the western and central belts of Ordovician volcanic rocks (R. Scott unpublished data). After extension, a further deformation pulse resulted in the Girilambone–Wagga terrane being thrust obliquely over the arc while undergoing southwards translation. This second pulse also resulted in deformation of Llandovery turbidites and the syn-deformational emplacement of granites.

In the strike-slip model, the Late Ordovician–Early Silurian 'Benambran' deformation in the Bendigo terrane occurred while that terrane was still 'moored' off West Antarctica. Deformation intensity decreased towards the ocean, with distal parts of the fan system overlain conformably by Silurian strata filling the Melbourne Trough in the Tabberabberan cycle.

Tabberabberan cycle

Jamming of the subduction zone by the Bega terrane at the end of the Late Ordovician (Fig. 16d) caused rollback of the proto-Pacific plate until a new arc and inferred west-dipping subduction zone were established in the Late Silurian, further east in the present New England Orogen. The Tabberabberan cycle thus represents convergent margin relationships in the New England Orogen and back-arc basin extension in the Lachlan Orogen behind the arc. Several previous authors have argued for an extensional setting for the Tabberabberan cycle (e.g. Powell 1983, who envisaged extension in a dextral transform setting; Cas 1983; Scheibner 1989; Glen 1992; Scheibner & Veevers 2000; Collins 2002).

The Lachlan and the North Queensland orogens were characterized by widespread rifting that was marked by the formation of rift/extensional basins and the emplacement of vast amounts of granite. In North Queensland, many of these granites were emplaced into Precambrian basement west of the North Queensland Orogen, separated from it by an extensional fault detachment (Blewett & Black 1998; P. Donchak pers. comm. Fig.; 9). Rifting in the Lachlan Orogen was terminated by the Mid Devonian Tabberabberan Orogeny: in the


(a)
Early and Middle Ordovician
AUSTRALIA
Thomson Orogen
Delamerian Orogen
Victoria Land
Ross Orogen
Transantarctic Mountains
Marie Byrd Land
W T
direction of spreading
Submarine fans
Legend
Macquarie Arc
volcaniclastic forearc basin & accretionary prism
Girilambone - Wagga Terrane
Bega Terrane
Bendigo Terrane
(b)
early Late Ordovician
seamount
subduction jammed
(c)
Late Ordovician
(d)
Late Ordovician to Early Silurian
Narooma Terrane
(e)
Early Silurian to Mid Devonian
Melbourne Trough
(f)
Cretaceous
Anakie
New England Orogen
Tasmania
Tuhua Orogen
NEW ZEALAND

North Queensland Orogen by a less clearly defined Devonian–Carboniferous deformation. The lack of a prominent Mid-Devonian orogeny is not surprising if this deformation was driven by strike-slip tectonics in Victoria (see below).

Rift basins of the Tabberabberan cycle in the Lachlan Orogen were discussed earlier. The large-scale crustal structure of the Tabberabberan cycle is still uncertain. Some of the shallowest basins (e.g. Cowra Trough) contain large amounts of volcanic rocks and intrusive granites that do not imply large extension, since the fill is only a few kilometres of generally shallow water to subaerial material (David & Glen unpublished data). Either the trough margins have been strongly overthrust or the heating is related to asymmetrical extension across the Lachlan Orogen (e.g. Lister *et al.* 1991).

One major question is how the different basement rocks responded to extension. The deformed Ordovician Macquarie Arc seems to have been pulled apart by extensional/oblique faults, with rifted parts forming basement to the deep-water Hill End Trough, the shallower Jemalong Trough and part of the Cowra Trough, as well as shallow-water flanking shelves (Glen *et al.* 2002; Vassallo *et al.* 2003). Unrifted parts occur as dispersed structural belts (Glen *et al.* 1998). Seismic reflection profiling suggests that these extensional faults are largely planar (Glen *et al.* 2002). However, the Tumut Trough, underlain by a complex basement of accreted Macquarie Arc juxtaposed against Ordovician turbidites and MORB-type volcanic rocks (Meffre & Glen unpublished data), formed by north–south transtension, and Stuart-Smith (1990) described a low-angle extensional detachment at the basement–cover interface.

Basement consisting of Ordovician turbidites also underwent widespread mid-Silurian to mid-Devonian subsidence. This led to deposition of deep-water troughs but also shallow-water (to subaerial) shelves, commonly containing largely felsic volcanic and volcaniclastic rocks with lesser amounts of basalts and andesites. A key difference is the presence of large granitic batholiths intruding these Ordovician turbidites. It appears that clues to extension mechanisms and geometries in the middle crust may come from study of the widespread granites, which reflect major transfer of material from the lower crust and upper mantle into an actively extending middle crust.

In the Lachlan Orogen, many authors noted that these voluminous granites were generated and emplaced in a back-arc extensional environment (e.g. Scheibner 1987; Fergusson 1992*a*; Glen 1992; Scheibner & Basden 1998). One-dimensional modelling by Zen (1995) indicated that S-type melts could be generated in thinning crust. He also suggested that I-type granites east of the I–S line might reflect deeper-level melting in less extended crust. Chappell & co-workers have maintained that granite generation was unrelated to subduction: only the easternmost suite of I-type granites possesses geochemistry resembling East Pacific-type cordilleran granites (Chappell 1984). Other I-type granites were derived from 500–600 Ma tonalitic igneous crust (Chappell & Stephens 1988; Williams & Chappell 1998). S-type (and I-type) granites are now thought to be sourced from Ordovician turbidites underthrust during the Benambran Orogeny (Glen 1992) and mixed in various ratios with a mafic basaltic component (Gray 1984) or with two components – a mantle-derived component and a component geochemically similar to Cambrian mafic volcanic rocks (Keay *et al.* 1997; Collins 1998). Detrital zircon

Fig. 16. The strike-slip model. (**a**) Early and Middle Ordovician showing convergent and strike-slip nature of plate boundary and distribution of Adaminaby superterrane as series of fans along the east Gondwana margin. The Macquarie Arc, opposite the convergent plate margin in north, and the Bendigo terrane were built on rifted pieces of Delamerian forearc and post-collisional (p.c.) volcanic rocks. Western Tasmania (WT) was built on the Delamerian Orogen, and was the site of platform sedimentation. If the Bega and Bendigo terranes were sourced from the Ross Orogen, they would lie on the Gondwana Plate. If they were sourced from the Mozambique Belts they would occupy a diffuse transform plate margin. (**b**) Movement of the Bega terrane away from the Gondwana margin and its northward translation is reflected by the Late Ordovician condensed black shale sequence. The onset of this translation was synchronous with a hiatus in volcanism in the Macquarie Arc (due to seamount impinging on the trench) and with inferred back-arc basin spreading in the Girilambone–Wagga terrane. (**c**), (**d**) The Bega terrane was translated north along the plate margin to lie outboard of the Macquarie Arc, where it blocked oblique subduction and was then accreted obliquely to the Macquarie Arc, which was driven west into the Girilambone–Wagga terrane. Bendigo terrane was deformed off the Antarctic sector of the Gondwana margin. (**e**) Northwards translation of the Bendigo terrane during fill of the Melbourne Zone from sources in the south, west and southwest until just before accretion when an eastern source became available. Accretion involved underthrusting of the Girilambone–Wagga terrane which moved south. (**f**) Cretaceous reconstruction after (Lawver & Gahagan 1994) showing final distribution of terranes.

work (Williams & Chappell 1998) confirms that the source need be no older than Early Ordovician, and the fertile nature of these rocks is confirmed by the recognition that older parts of the turbidite pile are more lithic and felspathic than the upper parts (M. Scott & O.Thomas, pers. comm.).

Melting to produce these Silurian–Devonian granites probably began during thrust thickening of Ordovician turbidites in the Benambran Orogeny (Fagan 1979; Pogson 1982; Glen 1992) and continued during crustal extension. Collins & Hobbs (2001) showed that input from mantle melts was also needed to generate granitic magmas. Mantle input is also supported by the association of intrusive gabbro to dioritic plutons and associated dykes around granites and also by the presence of mafic–intermediate enclaves (e.g. Gray 1984; Soesoo & Nicholls 1999; Collins & Hobbs 2001). The subduction signature of some Silurian–Devonian granites and volcanic rocks is ascribed to inheritance from Ordovician subduction (e.g. Watkins 1998b; Glen 1998), rather than indicating the roots of an arc (Soessoo *et al.* 1997; Collins & Hobbs 2001 see below), although geochemical modelling is needed to confirm this possibility. This argument becomes stronger when examined in the light of the change from Late Carboniferous convergence to Early Permian extension in the New England Orogen. Several authors cited in Part 2 of this paper have pointed out that the first melts in the Early Permian carry subduction signatures inherited from Carboniferous subduction (e.g. Brownlow & Arculus 1999; Caprarelli & Leitch 2001; Jenkins *et al.* 2002).

The Tabberabberan Orogeny at the end of the Tabberabberan cycle reflects accretion of the northward-shunting Bendigo terrane to Gondwana (Glen *et al.* 1992*b*; VandenBerg *et al.* 2000; Willman *et al.* 2002) (Fig. 16e). This terrane, structurally represented by the Southwestern subprovince, was inserted between the Delamerian Orogen on the west in the Late Siluarian–Middle Devonian by a combination of mild north–south shortening, coupled with strike-slip deformation and strong east–west shortening (Glen *et al.* 1992; Miller *et al.* 2001) and the Eastern and Central subprovinces of the Lachlan Orogen on the east in the Middle Devonian. As it moved north, it received a mixture of sandstone and shale from sources in the west and south: easterly sources only became apparent as the terrane approached the Gondwana margin in the late Early Devonian. The strike-slip model of VandenBerg *et al.* (2000) and Willman *et al.* (2002) has opposite dynamics to that of this paper, with Eastern Australia (Eastern and Central subprovinces) moving dextrally SSE rather than the Bendigo terrane moving to the north.

Other models

Plate tectonic models in the early 1970s envisaged only one subduction zone at the western margin of the proto-Pacific plate (e.g. Oversby 1971; Solomon & Griffiths 1972). Updates of these ideas were presented by several authors (e.g. Powell 1984*a*; Scheibner 1985; Coney 1992). Models invoking multiple subduction zones were also proposed, e.g. the multiple back-arc model of Scheibner (1973, but not later), the sequential forearc model of Crook (1980) and back-arc model of Collins & Vernon (1992). The last suggested two subduction zones behind the plate boundary: an arc related to east-dipping subduction along the boundary between the Central and Southwestern subprovinces at 435 Ma, and a west-dipping subduction zone at *c.* 400 Ma within the Southwestern subprovince to close the Melbourne Trough, interpreted as a back-arc basin. Gray and co-workers further developed the concept of multiple subduction zones, based on Ar–Ar dating in the Southwestern subprovince that showed that the oldest cooling ages were Ordovician and became younger (mid-Devonian) in the east (Gray *et al.* 1997; Foster *et al.* 1999; Foster & Gray 2000) (Fig. 6). This eastward younging in cooling ages was interpreted as a single 'diachronous' transgressive deformation extending over 50 million years (Gray *et al.* 1997). This concept of continuous deformation led Gray & Foster (1997: 880–881) to suggest that 'the time from what was previously defined as Benambran to Tabberabberan is redefined as one progressive orogenic episode that we now call the Lachlan Orogeny after Cas (1983)'. This prograding deformation was likened to that occurring in an accretionary wedge above a subduction complex, with migrating deformation related to 'subduction accretion during plate convergence' in an oceanic setting with three subduction zones active in the mid-Palaeozoic (Gray *et al.* 1997: 497). Fault vergence was thus regarded as antithetic to the dips of subduction zones. However, modelling by Keep (2003) showed that oceanward-verging thrusts could form without subduction of oceanic lithosphere. Whereas Gray & coworkers suggested that the Benambran Orogeny reflected progressive growth of an accretionary wedge, Collins & Hobbs (2001) suggested that the two phases identified above reflect two separate, synchronous subduction–acrcetion/magmatic arc complexes, with Early Silurian S-type granites forming the roots of magmatic arcs. Both are in contrast to the model of this paper.

Subduction zone 1 in the west (Gray & Foster 1997) was shallowly west-dipping and existed from 460 Ma to 420 Ma (Foster *et al.* 1999). No arc was associated with this zone. Cambrian–Ordovician turbidites in the Stawell and Bendigo structural zones, deformed into east-vergent structures, were interpreted as having formed in the associated accretionary prism that was advancing progressively eastward and closing a marginal basin (Foster *et al.* 1999). Subduction zone 2 was short lived (440–420 Ma, Foster *et al.* 1999), was east-dipping and lay at the boundary between the Melbourne and Tabberabbera zones (between the Central and Southwestern subprovinces). There is no arc associated with this subduction zone, although some syn- to post-tectonic granites in the Central subprovince have subduction-related signatures. Ordovician turbidites with southwest-verging folds (Fergusson 1987) in the Tabberabbera and Omeo zones were interpreted as accretionary prism rocks and reflections of a magmatic arc. Subduction zone 3 was the Gondwana–proto-Pacific plate boundary. The associated arc was the Macquarie Arc of Glen *et al.* (1998). The oceanic Narooma Terrane and surrounding Ordovician turbidites were interpreted as the related accretionary prism, after Miller & Gray (1996), although this was argued against by Glen *et al.* (2004).

The multiple subduction model was taken further by Soesoo *et al.* (1997), who suggested that subduction zone 1 persisted until the Middle Devonian (*c.* 380 Ma) and that the Tabberabbera zone was underlain by two opposing subduction zones – the east-dipping subduction zone 2 of Foster *et al.* (1999) and an extra west-dipping zone – from 420 to *c.* 380 Ma. The existence of this subduction zone is supported by the presence of *c.* 450 Ma blueschist knockers along the boundary between the two structural zones (Spaggiari *et al.* 2002) and by the subduction signatures of granitic rocks (Nicholls *et al.* 1996). The 'Lachlan Orogeny' was terminated by the late Early to Middle Devonian closure of the marginal basin (original width of *c.* 750 km, now the Melbourne Trough) and by docking of an island arc/forearc system to the east (Gray & Foster 1997).

Negative Nb and Ti anomalies in some Early Devonian granites have been used to infer the influence of subduction on mantle magma sources, suggesting that these tracts of Late Silurian and Early Devonian granites are the roots of coeval subduction-related arcs (Soesoo *et al.* 1997). In contrast, Middle and Late Devonian mafic rocks lack significant Nb anomalies and were interpreted as emplaced in continental–rift extensional settings (Soesoo & Nicholls 1999). Collins & Hobbs (2001) proposed the existence of two coeval subduction zones from 435 Ma to 425 Ma, with arcs reflected by the distribution of NW- and N-trending belts of largely S-type granites. The multiple subduction model was extended by Fergusson (2003) who argued for four subduction zones

In contrast to the multiple subduction models, Collins & Vernon (1994) suggested a model involving east–west sequential delamination of lithospheric slabs, with granitoids generated from basaltic underplating rather than from subduction melts as in the earlier model. VandenBerg *et al.* (2000) and Cayley *et al.* (2002) argued for a vice model, in which the deformation of the Southwestern subprovince was driven by compression between the Delamerian Orogen backstop in the west and a block of Proterozoic continental crust under the Melbourne Zone in the east, driven westwards by plate boundary forces. This 'rifted-off' piece of continental crust has been incorporated as an alternative into the multiple subduction model by Gray *et al.* (2003).

Subduction zones, however, require special conditions to form and do not turn on and off easily. Strike-slip 'transpressional' tectonics, coupled with models in which the subduction signatures of Silurian–Early Devonian granites are inherited from Ordovician subduction, are alternatives to the multiple subduction zone models. The problem with strike-slip tectonics is that it is based largely on palaeogeographical reconstructions, since kinematic evidence of hundreds of kilometres of strike-slip movement has not been established. However, it is argued that strike-slip deformation must be the norm in convergent margin orogens (e.g. Teyssier & Tikoff 1995), except in rare cases where convergence is head-on to a perfectly planar plate boundary, and that much of it can be achieved by summing up smaller (tens of kilometres) displacements on a number of smaller faults.

Concluding discussion

The Tasmanides of eastern Australia have been part of Gondwana, and then part of Pangaea before it began to break apart at *c.* 230 Ma. The Tasmanides faced the proto-Pacific Ocean following supercontinent break-up at *c.* 750 Ma.

There is no evidence of continent–continent collision, or that the Wilson Cycle of ocean opening and closing ever operated (Crawford *et al.* 2003*a*; Cawood 2002, 2005). Crook (1969) first pointed out this difference between Atlantic and Pacific geosynclines.

Most of the history of the Tasmanides records extension or rifting. With the exception of passive margin development during rifting of Rodinia in the Neoproterozoic, it is envisaged that all subsequent evolution of the Tasmanides occurred in a convergent margin setting along the proto-Pacific plate. Phases of extension and rifting are separated by deformation events that occupy only short intervals during the development of the Tasmanides (Fig. 4). They were probably more complicated than current models suggest, and probably diachronous along-strike.

It was suggested previously (Glen 1995) that the Lachlan Orogen, in particular, possessed many of the features of a convergent margin in which the rate of convergence of the two plates was less than the rate of subduction (cf. Royden & Burchfiel 1989; Royden 1993). Such an orogen is characterized by extension, lack of major collision and resultant high topography, and an absence of faults that exhume high-grade basement. Today, one would call this a retreating or accretionary orogen, defined by rollback of the proto-Pacific plate (Cawood 2002; Collins 2002). To some extent, this concept can be applied to the other parts of the Tasmanides (e.g. Cawood 2002). However, there is little evidence for rollback of the proto-Pacific Plate in the North Queensland Orogen, and the plate boundary in the New England Orogen was fixed in position effectively from the Late Devonian to the Late Permian. This evolutionary model for the Tasmanides – accretionary in nature, with distinct, short periods of advance or accretion/collision events (e.g. Collins 2002; Crawford 2003*a*) – differs from that of Gray and co-workers for the Lachlan Orogen, in which most of the deformation of the Lachlan Orogen occurred in propagating accretionary wedges, with deformation reflecting times when wedges were thickened by internal deformation. Despite occupying small intervals of time, some orogeny (at least) seems to have been complex and either protracted or multiphase. The Delamerian Orogeny in the outboard parts of the Delamerian Orogen (western Tasmania and western Victoria) consisted of several phases, but seems to be more simple than the same orogeny in the inboard part of the orogen, where dating of foliated granites suggest that it lasted for 26 million years. The Benambran Orogeny is another protracted or multiphase event lasting *c.* 10 million years.

In accretionary orogens, what produces deformations? There are several possibilities.

(a) Changes in plate motions. The evolution of modern SW Pacific tectonics is generally seen as due to responses to changes in plate motions (e.g. Crawford *et al.* 2003*a*). A possible candidate in the Tasmanides is the inferred Carboniferous accretion of the New England Orogen to the Lachlan Orogen as the driver for the Kanimblan Orogeny, for which no other cause is clear. Veevers (2000*b*) pointed out that this deformation could have been a far-field result of the collision between Gondwana and Laurussia, although it seems to be a little older than the main collision phase in the Variscan belt (*c.* 340 Ma cf. 330–320 Ma, Veevers 2000*b*).

(b) Change in plate boundary dynamics or in coupling, from retreating to advancing, possibly in response to changes in spreading rates in (a). Carboniferous–Triassic deformation and extension in the Hunter–Bowen supercycle orogen were attributed to this mechanism by Jenkins *et al.* (2002). On a larger scale, differences between the Delamerian and New England orogens, on one hand, and the Lachlan Orogen, on the other, might reflect formation along largely advancing, as opposed to largely retreating, parts of the plate margin. Both the Delamerian and New England orogens, despite their differences, are classical orogens in the sense of having highly deformed and metamorphosed internal parts, containing accreted complexes and less deformed external parts consisting of fold–thrust belts. The Lachlan Orogen is more of a retreating orogen with deformations that did not produce high topography (Glen 1992).

(c) Collision between the Gondwana margin and a collider such as a large seamount or submarine plateau or island arc on the proto-Pacific plate. Examples in the Tasmanides include the arc–continent collision that produced the Delamerian Orogeny and, perhaps, also the accretion of the Gympie Terrane with Gondwana to produce the Triassic part of the Hunter–Bowen Orogen.

(d) Shuffling of terranes along-strike in a 'transpressional' regime that also involved shortening. Examples include the Benambran Orogeny as a response to shuffling of the Bega terrane to a location outboard of the Macquarie Arc, located on the Gondwana plate, and its resulting accretion, which drove the arc into its back-arc basin. The Tabberabberan Orogeny, which developed as a result of accretion of the Bendigo terrane with other parts of the Lachlan Orogen, is another example. The end-Silurian Bindi deformation seems to be a smaller-scale version of this, reflecting strike-slip-driven movements on linked faults.

There are several other key features that can be distilled from the synthesis of the Tasmanides.

(1) There are significant differences in the evolution of the Lachlan and North Queensland orogens, despite the general recognition of elements of the Lachlan supercycle in the North Queensland Orogen, indicating that both underwent Ordovician convergence followed by Silurian to Middle (or Late) Devonian extension. Their subsequent evolution differs: there is no Kanimblan cycle, represented by major foreland-style fluviatile deposits, in the North Queensland Orogen.

The primary difference between the two orogens is that the North Queensland Orogen was developed on continental crust of the Delamerian cycle, whereas the Lachlan Orogen formed on oceanic crust east of the Delamerian margin. The second difference flows from this – the North Queensland Orogen lies immediately east of the inboard boundary of the Tasmanides, whereas the Lachlan Orogen formed well to the east of the Tasmanides margin.

These differences suggest that the rapid rollback of the Gondwana–proto-Pacific plate boundary in southeastern Australia – after the Delamerian and Benambran orogenies – was not mirrored in North Queensland, where the plate boundary was more or less anchored in time from the Cambrian through to the Triassic. This lack of major back-arc spreading is, in turn, reflected in the continental margin nature of inferred arc(s) rather than the intra-oceanic nature of the Lachlan arcs. The evolution of the North Queensland Orogen is, thus, to a limited extent, more akin to the evolution of Palaeozoic South America (e.g. Ramos 1988; Rapela *et al.* 1998). The morphology of the proto-Pacific margin of Gondwana must have reflected this disparity, either by segmentation or by variation of spreading.

(2) The relationship between New England Orogen and orogens to the west is uncertain. Although there is strong continuity and parallel development between the northern and southern parts of the New England Orogen, there are striking differences between the New England Orogen and the Lachlan, Thomson and North Queensland orogens to the west, all of which occupied back-arc positions from the Middle Silurian to the Triassic.

Similarities between the Cambrian to Ordovician geology of the Lachlan Orogen and that of Cambrian–Ordovician fault-bounded blocks along the Peel–Manning Fault System in the southern New England Orogen were pointed out by Cawood & Leitch (1987) and Glen & Scheibner (1993). These similarities are interpreted here in terms of the Lachlan rocks being translated north in the Llandovery, to form substrate to the developing New England Orogen, and exhumed along the Peel–Manning Fault System. (Alternatively, Aitchison & Ireland (1995) suggested the New England Orogen overthrust the Lachlan Orogen in the Late Carboniferous.)

In the Silurian and Devonian, the two orogens had very different histories, despite the southern part of the New England Orogen lying outboard of the Lachlan Orogen after restoration of the Permian oroclines. This contrast is most evident in the Late Devonian, when fluviatile siliciclastic sediments extended from the Delamerian Orogen to the present NSW south coast, and when the western part of the New England Orogen was the site of a continental arc. The presence of volcanic detritus in the Lambie facies east of the Hill End Trough (Powell *et al.* 1984) suggests proximity, as does the presence of Lambie-facies type clasts in the forearc basin of the New England Orogen in the Late Viséan–Namurian (Cawood & Leitch 1985). If the New England Orogen was allochthonous (Klootwijk 2002), it was approaching the Lachlan Orogen in the Late Devonian, and docked in the Early Carboniferous, at about the time of the Kanimblan Orogeny.

The lack of a Lambie facies in the Thomson and North Queensland orogens reduces this contrast with the northern New England Orogen. Both the Thomson and North Queensland orogens have undergone extension, contraction and granite emplacement, extending into the Triassic, all consistent with their being back-arc systems to the New England Orogen.

Several questions still remained to be answered, not the least of which is more closely tying granite genesis into upper crustal histories. Another question not touched on here is how to reactivate the orogen-normal and orogen-oblique structures that seem to persist through these cycles and through the change from oceanic to continental crust.

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