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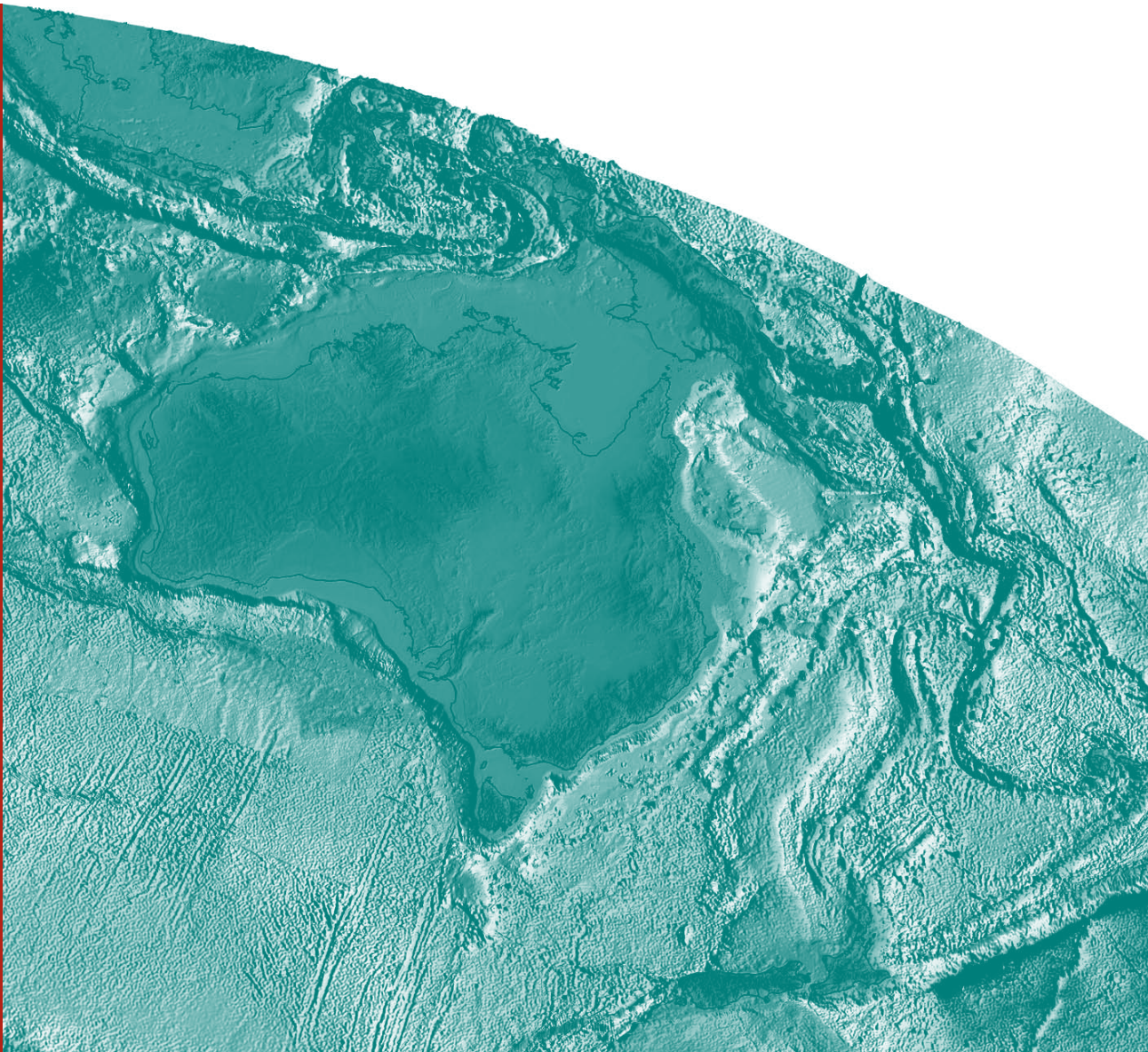
Geodynamic Synthesis of the Phanerozoic of Eastern Australia and Implications for Metallogeny

D.C. Champion, N. Kositcin, D.L. Huston, E. Mathews and C. Brown

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GEOSCIENCE AUSTRALIA
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by

D. C. Champion¹, N. Kositsin¹, D.L. Huston¹, E. Mathews¹, and C. Brown¹



Australian Government
Geoscience Australia

1. Onshore Energy and Minerals Division, Geoscience Australia, GPO Box 378, Canberra ACT 2601

Department of Resources, Energy and Tourism

Minister for Resources and Energy: The Hon. Martin Ferguson, AM MP

Secretary: John Pierce

Geoscience Australia

Chief Executive Officer: Dr Neil Williams PSM

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Introduction

by DC Champion and N Kositsin

This report presents the results of a geodynamic synthesis of the Phanerozoic of Eastern Australia. This was undertaken with the dual aims: (1) to better understand the tectonic and geodynamic setting of existing mineral deposits within eastern Australia, and (2) to provide a predictive capability, within the synthesised geodynamic framework, for extending potential regions of known mineralisation and for delineating regions with potential for new mineral styles and commodities. The report combines the Mineral Systems approach of Wyborn et al. (1997) with the ‘Five Questions’ methodology adopted by the pmd*CRC (<http://www.pmdcrc.com.au/RESprograms.html>; Barnicoat, 2007, 2008). It is clearly targeted at the first of the ‘Five Questions’, namely, constraining and understanding the regional and local geodynamic environment as the first step in delineating mineral systems. To achieve this we have synthesised geological and metallogenic data on a regional, largely orogenic, basis. This was undertaken to identify geological and metallogenic events and geodynamic cycles, and to produce regional geological syntheses and accompanying time-space-event plots. These regional syntheses were used to produce an interpreted geological, metallogenic and geodynamic synthesis of eastern Australia. This new synthesis provided the geodynamic framework to constrain known mineralisation and allow a predictive capability for potential new mineralisation. Outputs are delivered in three parts. Geological summaries and time-space-event plots are presented in [Section 1](#). Our interpreted geological and geodynamic synthesis in [Section 2](#), and associated metallogeny (known and predicted) for eastern Australia in [Section 3](#).

This report has concentrated on the Tasmanides or Tasman Orogen or Tasman Orogenic Belt (Scheibner and Veevers, 2000; Veevers, 2000, 2004; Cawood, 2005; Glen, 2005) of eastern Australia ([Fig. 1a](#); see below). This essentially corresponds to the Palaeozoic and early Mesozoic rocks east of the old Gondwanan (Delamerian) continental margin.

Nomenclature and terminology

We use the term ‘*orogen*’ (e.g., Lachlan Orogen) simply to designate an orogenic province. It is used in a similar sense to the historical but incorrect use of fold belt (e.g., Lachlan Fold Belt). ‘*Orogeny*’ on the other hand refers to an orogenic event, typically accompanied by deformation, metamorphism and magmatism (e.g., Delamerian Orogeny). A number of major orogenies have occurred within the eastern Australian orogens. Tectonic cycles are used in the ‘Wilson cycle’ sense to record the cycle starting post-previous orogeny, typically involving renewed extension and ended by major orogeny - the name of the orogeny is used for the cycle (as adopted by Glen, 2005), e.g., Benambran Cycle records the period from post-Delamerian Orogeny to the Benambran Orogeny. The cycle approach has been used to simplify and attempt to unify eastern Australia in a tectonic framework. There are, however, a number of potential difficulties with this approach. Firstly, it is evident that the timing(s) of the terminal orogeny in each cycle may be slightly different between orogens and even within orogens (e.g., [Figure 13](#)). Although some of this in part reflects poor age constraints, it would appear that it is, in part, real, and reflects the multiple stages of long-lived and possibly diachronous orogenies, such as the Benambran Orogeny in Victoria (e.g., VandenBerg et al., 2000). Additional concerns include orogeny names for multiple events, the best example being the Hunter-Bowen Orogeny which extends some 35 million years (e.g., ~265 to ~230 Ma; Holcombe et al., 1997b; Korsch et al., in press b). Whether this should be considered a tectonic cycle is a debatable question. Thirdly, within each cycle, there are additional, usually minor orogenies, such as the Bindian Orogeny in the Tabberabberan Cycle (e.g., Gray, 1997; Glen, 2005). There are also problems with regional correlations. For example, the multiple orogenies of the Alice Springs Orogeny broadly coincide with the major orogenies of eastern Australia (e.g., Benambran, Tabberabberan, Kanimblan) – it becomes difficult, and perhaps futile, to equate orogenies in regions which were potentially affected by both Alice Springs and eastern Australian orogenies (e.g., Koonenberry Belt, western Thomson Orogen and overlying basins; e.g., [Fig. 20](#)). In north Queensland, it appears that there is a slight timing difference between the Kanimblan and Alice Springs (3) orogenies, and both names have been used, although the Alice Springs Orogeny there forms part of the Kanimblan or Hunter Bowen Cycle.

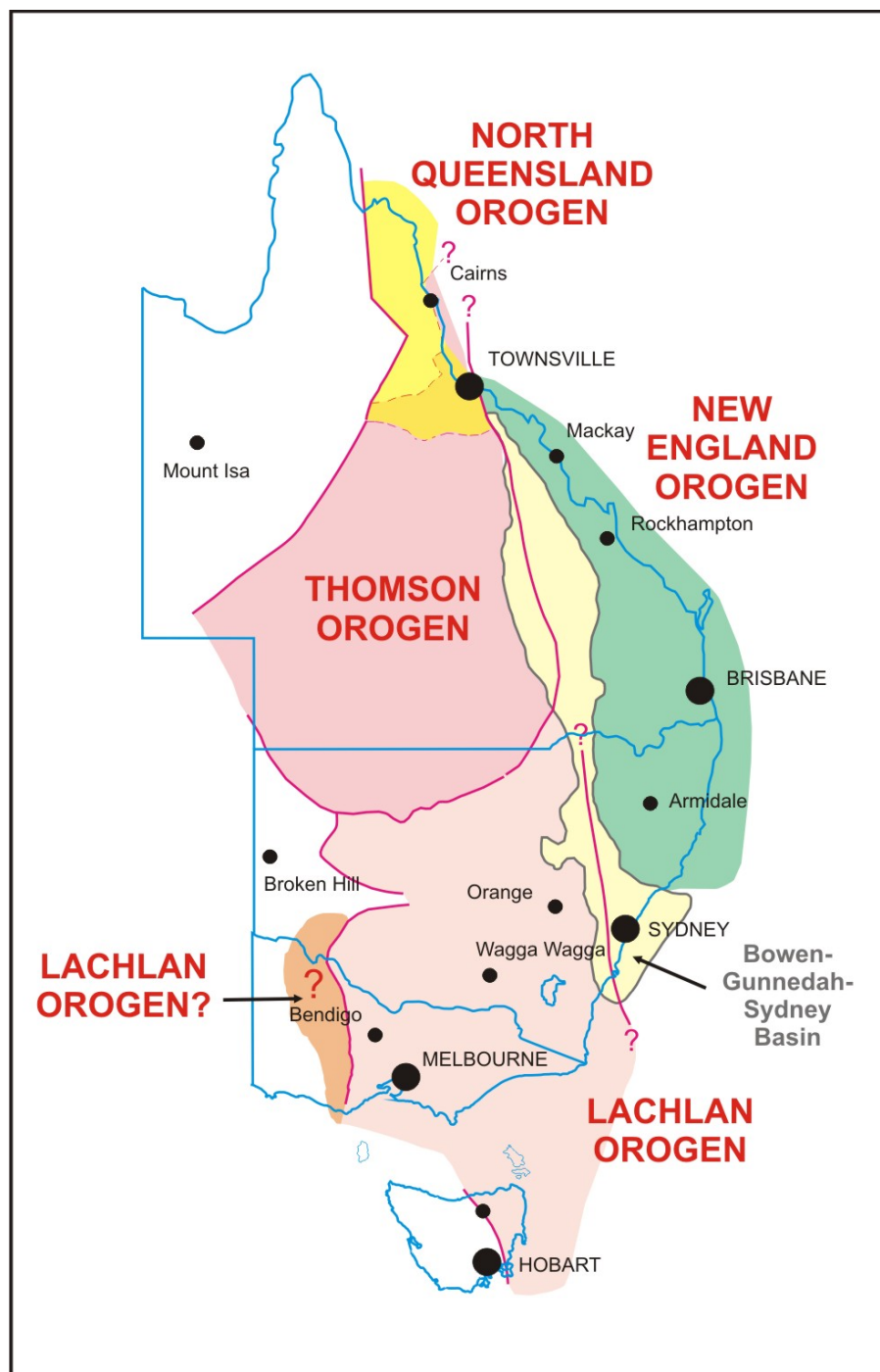


Figure 1a. Location map showing the distribution of eastern Australian orogens covered in this report. Orogen names follow Glen (2005). Orogen boundaries from Glen (2005), VandenBerg et al. (2000), Seymour and Calver (1995), Bain and Draper (1997), and unpublished GA-GSQ Nd isotope data for the eastern Thomson Orogen. The Thomson Orogen boundary has been extended to the north to include the Cape River and Barnard Provinces of Bain and Draper (1997). The Cape River Province is also included in the North Queensland Orogen, given its uncertain parentage. The Bowen-Gunnedah-Sydney Basin outline is from Geoscience Australia's 'Basins' national data set. The boundary of the Lachlan and Delamerian orogens is problematical and is either the western margin of the Bendigo or the Stawell zones (see Glen, 2005). We show it as a transitional zone.

Eastern Australia Orogenic Zones and tectonic cycles – definitions

The geology and tectonic development of the eastern Australia, particularly the Phanerozoic component - the Tasmanides - has been the focus of numerous past and continuing studies. This has resulted in a voluminous literature including numerous orogen-based or more regional reviews (e.g., Murray, 1986, 1997; Murray et al., 1987; Coney, 1992; Seymour and Calver, 1995; Bain and Draper, 1997; Gray, 1997; Gray and Foster, 1997; Gray et al., 1997; Scheibner and Basden, 1998; Foster and Gray, 2000; VandenBerg et al., 2000; 2004; Veevers, 2000; Cawood, 2003, 2005; Crawford et al., 2003a; Glen, 2004). The focus of this research has had two important outcomes. The first has been the recognition that eastern Australia, or more specifically the Tasmanides, can be broadly subdivided into a relatively small number of, largely Palaeozoic (and Mesozoic), complex and in part composite, orogens - Lachlan, Thomson, New England and North Queensland. These are located east of the more contiguous Proterozoic, and older, Australian continent that corresponds to the old Gondwana margin (e.g., Gray, 1997; Cawood, 2005). The boundary between the Tasman Orogenic Belt (TOB) and the post-Rodinian Gondwana margin is problematical. This corresponds to the old notion of the 'Tasman Line'. As summarised by Direen and Crawford (2003), apart from areas such as northern Queensland, where the Tasman boundary is clear, for most of its extent the exact location of the Tasman Line is enigmatic, and numerous variants of the line have been suggested. This is perhaps not surprising given the potential complexity of old craton boundaries and subsequent reworking, and both Direen and Crawford (2003) and Glen (2005) suggested the term was essentially meaningless, particularly in southern Australia. Regardless of this, it is clear that a boundary, albeit transitional, exists. Where this actually is depends in part on the definitions adopted (e.g., see discussion by Glen, 2005). For the purpose of this report, we have taken the middle ground. We have largely concentrated on those rocks east of the Delamerian Orogen, and the states these occur in, that is, Queensland, New South Wales, Victoria and Tasmania (Fig. 1a). We have, however, also taken in consideration Rodinian break-up and Delamerian Orogeny geology, but largely only where it occurs in the eastern States. We have not considered in any detail Tasmanide geology in other states, though refer to it where it is of significance, in particular the Delamerian Orogen in South Australia. Cawood (2005) has grouped the Tasman Orogenic Zone and the Delamerian Orogen, and its continuation throughout Gondwana, into the Terra Australis Orogen.

The definition of individual orogens and the subdivisions of these (zones, elements, regions, terranes, superterrane, provinces) are discussed in detail at the beginning of each section. We have, as far as applicable, followed the boundaries used by the relevant state surveys (e.g., Seymour and Calver, 1995; VandenBerg et al., 2000; Glen and co-workers, e.g., Glen (2005), Bain and Draper, 1997; and www.dme.qld.gov.au).

The second important outcome is captured in the time-space-event plots produced as part of this work and that is the recognition that there are broadly contemporaneous orogenic events recorded in all the orogens. This has been recognised before (e.g., the 'stages' recognised by Scheibner and Basden (1998) and Korsch and Harrington (1981) in NSW and the NEO, respectively) and has been used (e.g., Glen, 2005) to define tectonic cycles for all the Tasmanides. There is some divergence after the Benambran Orogeny, when the New England Orogen has slightly different cycles from the rest of mainland Australia. Although there are some potential difficulties with this approach, as outlined earlier, it is very useful in providing a methodology that allows unification of eastern Australia in a tectonic framework. We have followed this approach and, hence, the geological syntheses for regions and for eastern Australia presented here, have been documented on the basis of the Delamerian, Benambran, Tabberabberan, Kanimblan and Hunter-Bowen Tectonic Cycles. Potential difficulties as outlined earlier (e.g., different ages, multiple events) are discussed in each section for the relevant regions and cycles.

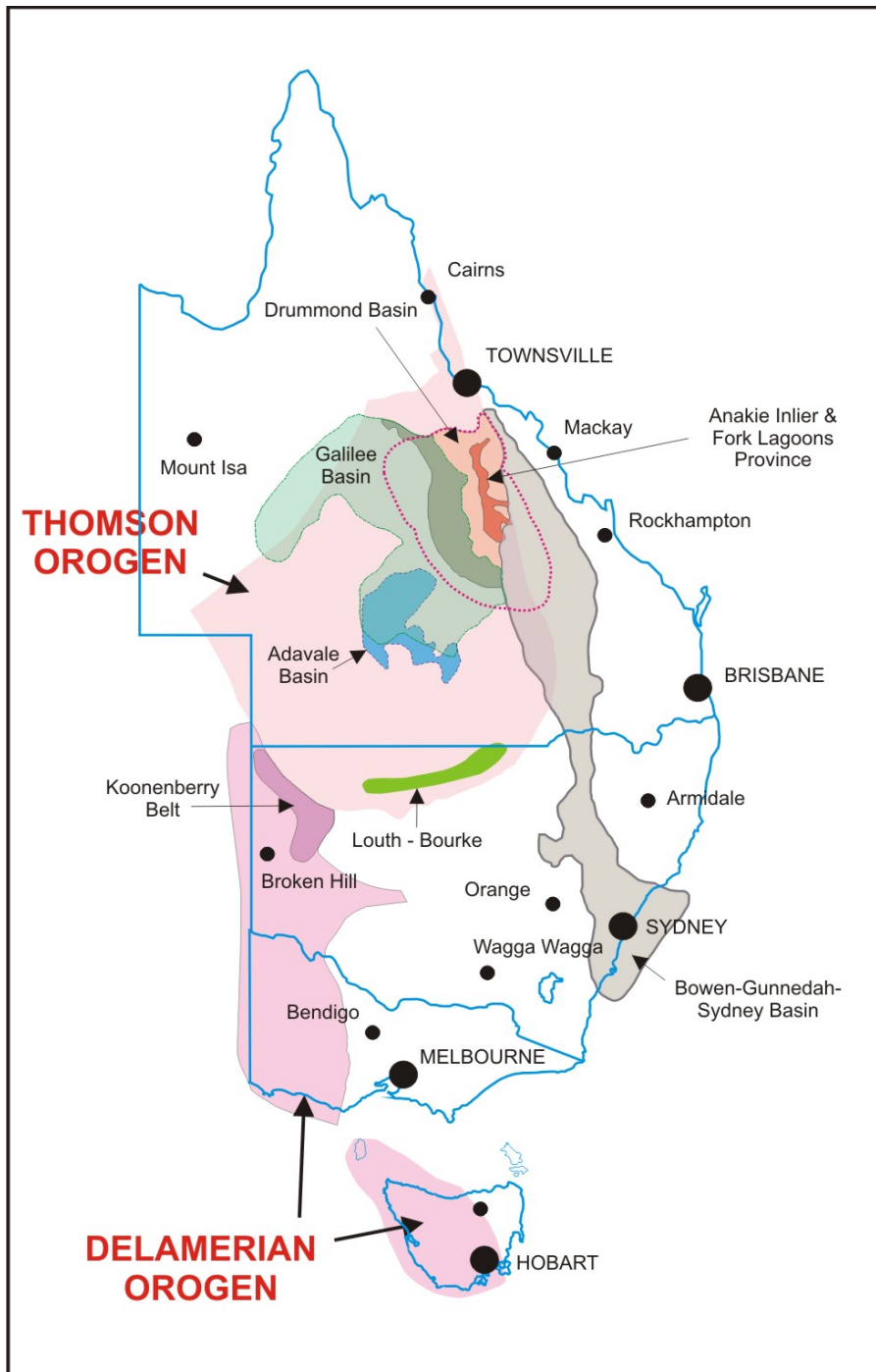


Figure 1b. Location map showing the distribution of eastern Australian orogens, basins and other regions. Orogen names and boundaries follow Glen (2005), Bain and Draper (1997), and unpublished GA-GSQ Nd isotope data for the eastern Thomson Orogen. The boundary of the Thomson Orogen is extended north to include the Cape River and Barnard provinces of Bain and Draper (1997). The Koonenberry Belt boundaries are after Gilmore et al. (2007). The Adavale, Galilee, Bowen-Gunnedah-Sydney, and Surat Basin outlines are from Geoscience Australia's 'Basins' national data set. Drummond Basin and Anakie Inlier boundaries from 'Geology of Queensland' (www.dme.qld.gov.au/mines/projects.cfm).

Acknowledgements

The report was greatly assisted by detailed discussions on state geology and metallogeny, as well as reviews by, personnel of the Queensland, New South Wales, Victoria and Tasmanian Geological Surveys. We particularly wish to thank C. Murray, J. Draper, L. Hutton, I. Withnall (Geological Survey of Queensland), R. Glen, and P. Blevin (Geological Survey of New South Wales), C. Willman (Geoscience Victoria), R. Bottrill, C. Calver, G. Green, M. McClennahan, D. Seymour and J. Taheri (Mineral Resources Tasmania). We also thank W. Collins (James Cook University) and R. Korsch (Geoscience Australia), who along with R. Glen (Geological Survey of New South Wales) gave invited presentations at Geoscience Australia on their views of Tasmanides Geology. We also acknowledge the 2008 Tasmanides Workshop, Melbourne – organized by Geoscience Victoria – and the speakers involved who summarized various aspects of Tasmanides Geology. We thank and acknowledge Ralph Bottrill (Mineral Resources Tasmania) who provided contributions, to [section 3](#), on aspects of Tasmanian metallogeny. We also thank Russell Korsch, Simon Van Der Wielen and Subhash Jaireth who provided detailed internal reviews at Geoscience Australia.

Section 1: Geological and geodynamic syntheses for orogens and regions of eastern Australia

by N Kositsin, DC Champion, E Mathews and C Brown

1.1. Lachlan Orogen and related margins

by DC Champion and N Kositsin

Introduction

The Lachlan Orogen, as used here, follows the general usage of numerous authors (e.g., Seymour and Calver, 1995, 1998; Scheibner and Basden, 1996; Gray, 1997; Gray and Foster, 1997, 2004; VandenBerg et al., 2000; Crawford et al., 2003b; Glen, 2005; Glen et al., 2007b), as shown in Figure 1a. The Orogen occurs within the central and eastern parts of New South Wales, Victoria and northeastern Tasmania (Fig. 1a). It is bound to the west by the Delamerian Orogen, to the north by the Thomson Orogen, and to the east by young oceanic crust or the New England Orogen. The Lachlan Orogen is traditionally interpreted as those regions not having undergone the Delamerian Orogeny (e.g., Glen, 2005). Included within the Orogen, however, are rocks which may be underlain by Delamerian crust, e.g., the Selwyn Block of Cayley et al. (2002; see below). The Delamerian-Lachlan boundary is poorly defined in western New South Wales, largely because of younger cover rocks. Similarly, some conjecture concerns the western boundary of the Orogen in Victoria. This is traditionally taken as the Moyston Fault on the western side of the Stawell Zone (e.g., VandenBerg et al., 2000; Crawford et al., 2003b). Recent work by Miller and co-workers (e.g., Miller et al., 2006) has shown that rocks of the western Stawell Zone have experienced Delamerian-aged deformation, indicating they are part of the Delamerian Orogen. We have taken the middle ground and show the Stawell Zone as a transitional boundary (Figs 1, 2) between the Lachlan and Delamerian Orogens.

The Lachlan Orogen itself has been subdivided in a number of ways, usually largely based on state lines. The geological surveys of both Victoria and Tasmania have each subdivided their state (and the Lachlan Orogen component) into geological regions – called zones in Victoria (VandenBerg et al., 2000; Fig. 3) and elements in Tasmania (Seymour and Calver, 1995, 1998; Fig. 4). The Lachlan Orogen within each state has also been subdivided into terranes and superterranes. These include the Whitelaw and Benambra Terranes in Victoria (e.g., VandenBerg et al., 2000; Fig. 3), and the Bega, Narooma, Girilambone-Wagga and Macquarie Arc terranes in NSW (e.g., Glen, 2005; Glen et al., 2007b; Fig. 2). Unfortunately, there is no agreement with terrane nomenclature and both the Girilambone-Wagga and Bega Terranes of New South Wales (Glen, 2005) contain parts of the Benambra Terrane of VandenBerg et al. (2000). Glen and co-workers also introduced the Adaminaby Superterrane (Fig. 2), for those regions of the Lachlan with Ordovician turbidites, consisting of the Bega, and Girilambone-Wagga terranes in NSW and the Bendigo, Tabberabbera and Omeo zones in Victoria (parts of VandenBerg et al.'s (2000) Whitelaw and Benambra Terranes). We refer to most of these zones and terranes (Figs 2, 3, 4) specifically when discussing relevant state geology. We, however, prefer the simpler terminology, also used by Gray (1997), VandenBerg et al. (2000), Gray et al. (2003), Glen (2005) of Western, Central and Eastern Lachlan (Fig. 2). The Western Lachlan – largely synonymous with the Whitelaw Terrane (VandenBerg et al., 2000; Figs 2, 3) is largely known from Victoria, but does extend into New South Wales (Glen, 2005). The Central and Eastern Lachlan occur in both Victoria and New South Wales (Fig. 2), though the majority of the Eastern Lachlan lies in New South Wales. Tasmania is more problematical. We follow VandenBerg et al. (2000) and Cayley et al. (2002) in adopting the Selwyn Block model (see below) which effectively links western Tasmania with the Melbourne Zone (Figs 1, 2, 4). In this scenario northeastern Tasmania is also included with the Melbourne Zone (Western Lachlan), though it shares similarities with both the Melbourne Zone and the Central Lachlan (e.g., Reed, 2001).

As outlined by Cayley et al. (2002), rocks of the Melbourne Zone probably formed upon Delamerian Orogen crust. Although this is not universally accepted (e.g., Gray and Foster, 2004; Spaggiari et al. 2004),

it indicates that the Melbourne Zone, like the Stawell Zone, can be thought of as transitional Delamerian-Lachlan. In the following discussion, although focused on the Lachlan Orogen, parts of the Delamerian Orogen are also included, particularly western Victoria, and Western Tasmania (including King Island). The Delamerian in South Australia, especially in the Stansbury Basin, is also referred to where necessary (see Fig. 8). The Koonenberry region and Warburton Basin is included with the Thomson Orogen discussion. Numerous significant reviews have been presented on the Lachlan Orogen, both on a state (e.g., Scheibner and Basden, 1996; Gray, 1997; VandenBerg et al. 2000; Birch, 2003; Seymour and Calver, 1995) and orogen basis (e.g., Gray and Foster, 2004; Glen, 2005) level. As a consequence although the general distribution of rock types is described, most attention is focused on identifying crustal blocks and interpreted geodynamic environments. Similarly, numerous time-space plots for the majority of the Lachlan, on a state or orogen-wide basis, have been presented previously (e.g., VandenBerg et al., 2000; Birch, 2003; Glen, 2005; Pogson and Glen, 2006; Seymour and Calver, 1998). Summaries of these are presented here (Figs 5, 6, 7), largely based on those previously published with minor modifications where required.

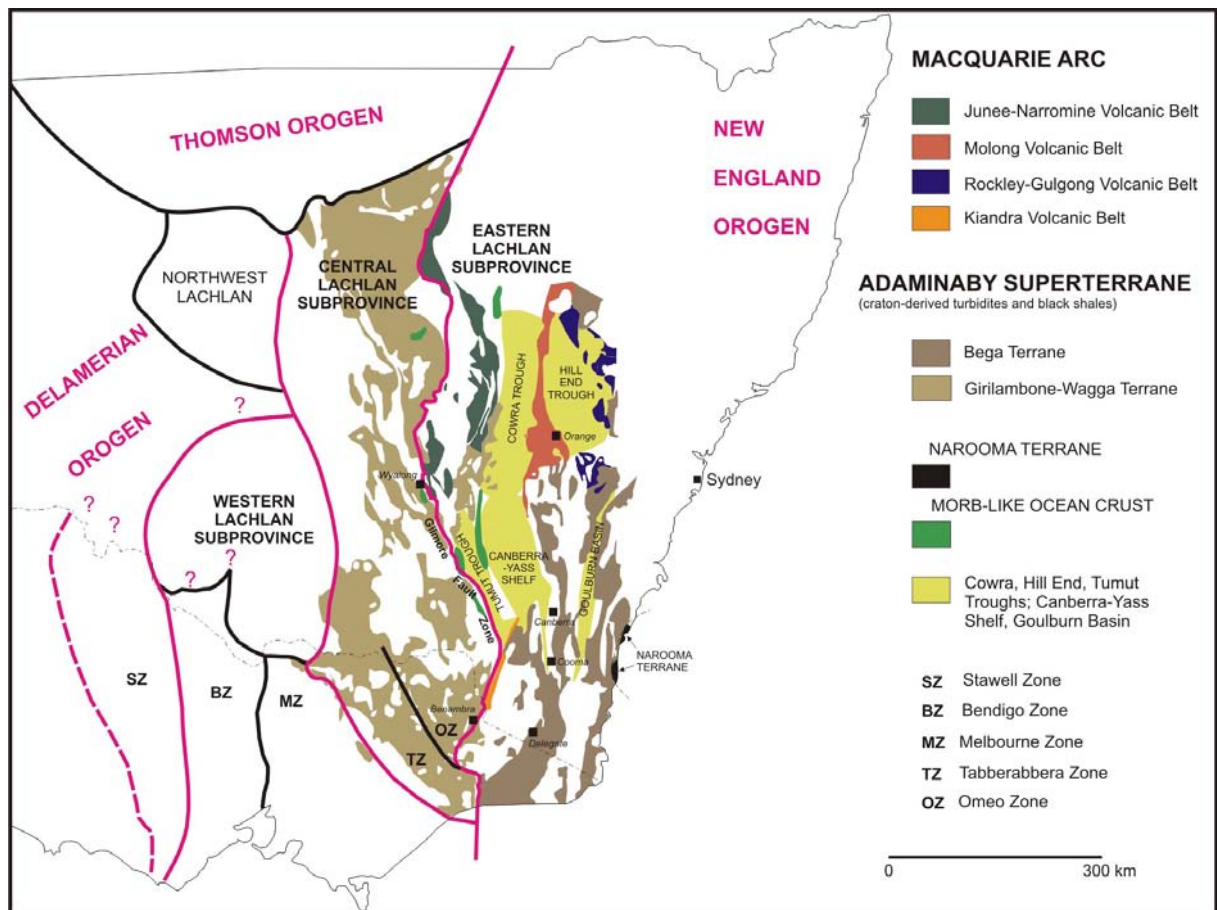


Figure 2. Zones, terranes and Lachlan subprovinces of New South Wales and Victoria. Figure modified after Fergusson (2003), Gray and Foster (2004), Glen (2005), Crawford et al. (2007a) and Glen et al. (2007d). Note in the definition of Glen et al. (2007b) the Bendigo Zone is also included in the Adaminaby Superterrane. Victoria zones follow nomenclature of VandenBerg et al. (2000). The Stawell Zone is considered a transitional zone with both Delamerian and Lachlan Orogen rocks, and it is treated as part of the Western Lachlan in this report.

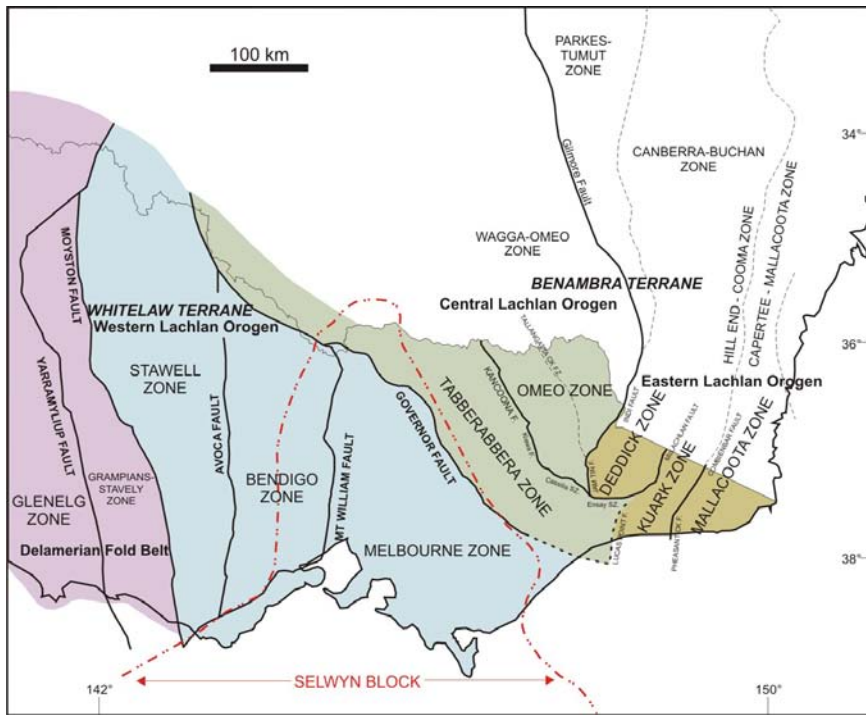


Figure 3. Geological orogens, terranes and zones of Victoria. Modified from VandenBerg et al. (2000). The Selwyn Block delineates the area thought to have Proterozoic basement underlying the Melbourne Zone (e.g., Cayley et al., 2002). The Whitelaw Terrane is largely synonymous with the Western Lachlan Subprovince (Fig. 2), though the location of the northern boundary is uncertain.

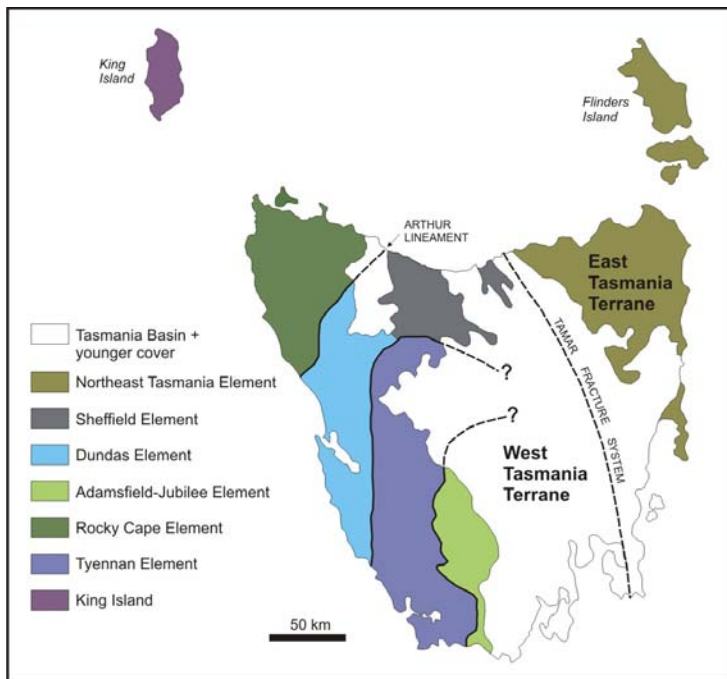


Figure 4. Tectonic elements of Tasmania. Figure modified from Seymour and Calver (1995, 1998). Eastern and Western Tasmania terminology adopted from Spaggiari et al. (2003). The correlation of Tasmania with mainland Australia is contentious. In the Selwyn Block model of Cayley et al. (2002), Tasmania is related to the Melbourne Zone (Western Lachlan; Figs 2, 3). East Tasmania is thought to correlate with either the Melbourne Zone or the Tabberabbera Zone (e.g., Reed, 2001).

Geodynamic Synthesis of the Phanerozoic of eastern Australia and Implications for metallogeny

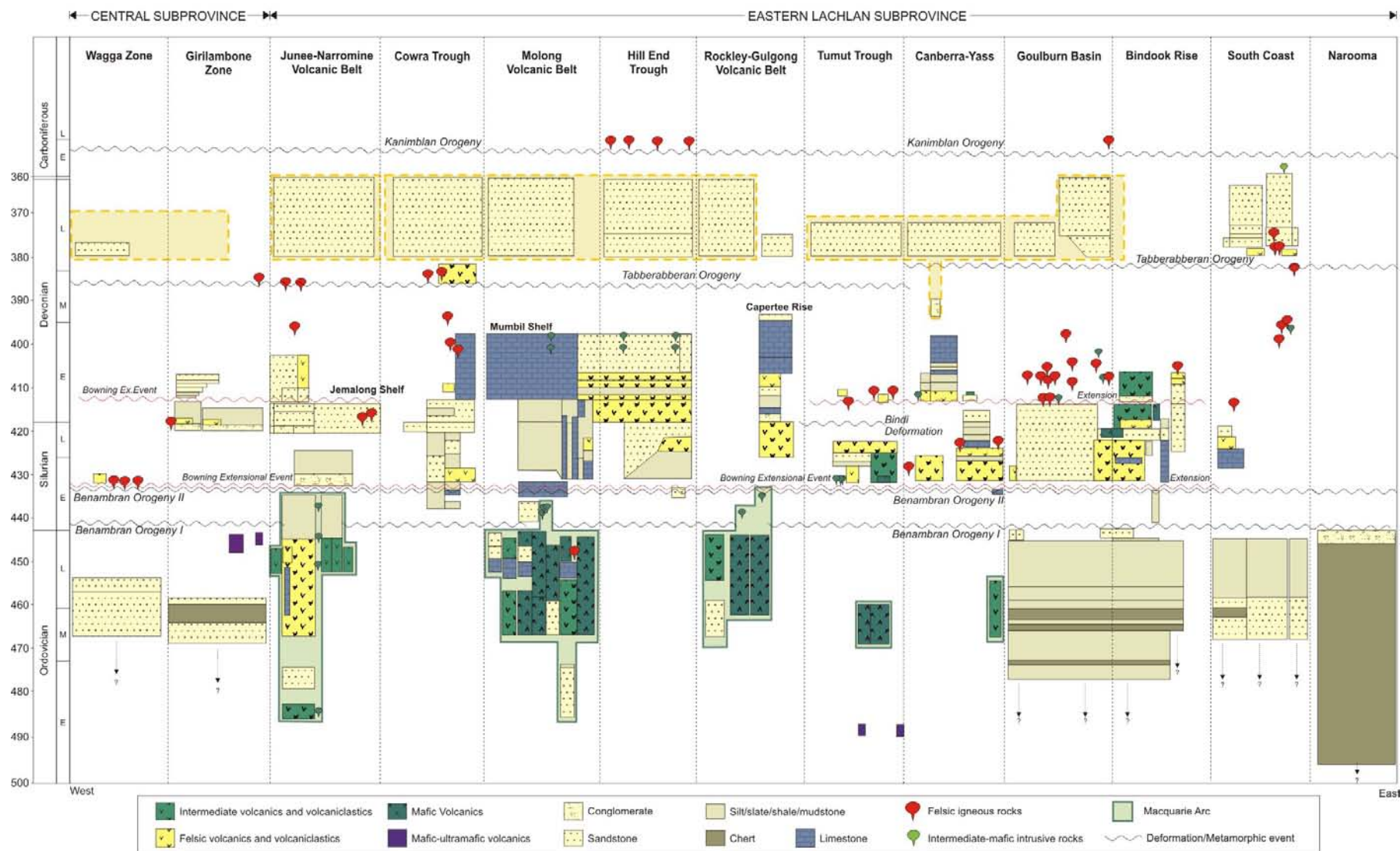


Figure 5. Time-space plot for the New South Wales part of the Lachlan Orogen. Figure modified from, and based largely on, the time-space plots in the East Lachlan Orogen GIS (Pogson and Glen, 2006).

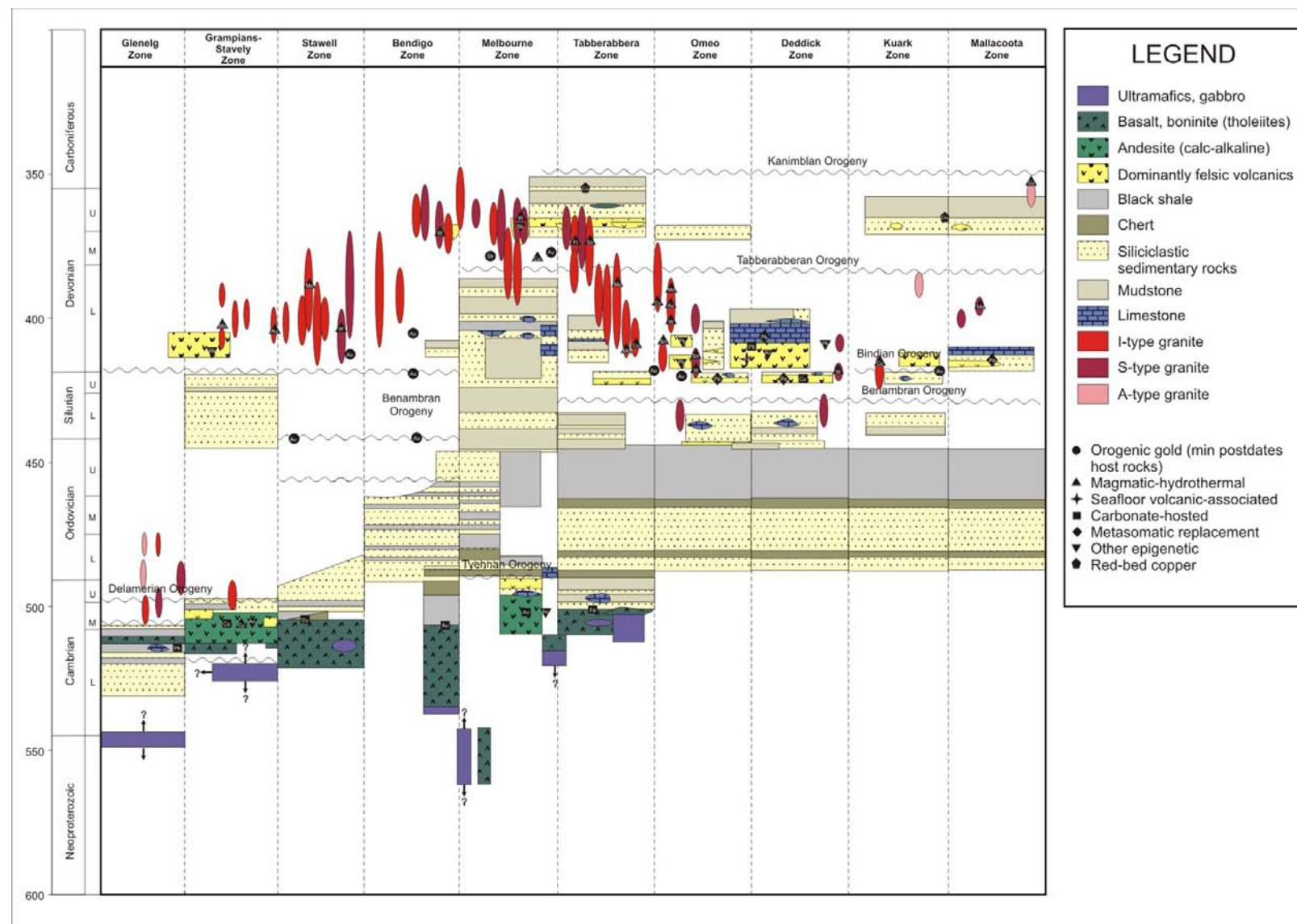


Figure 6. Time-space plot for Victoria. Modified from, and based largely on, the time-space plot of VandenBerg et al. (2000). Refer to [Figure 3](#) for location of zones.

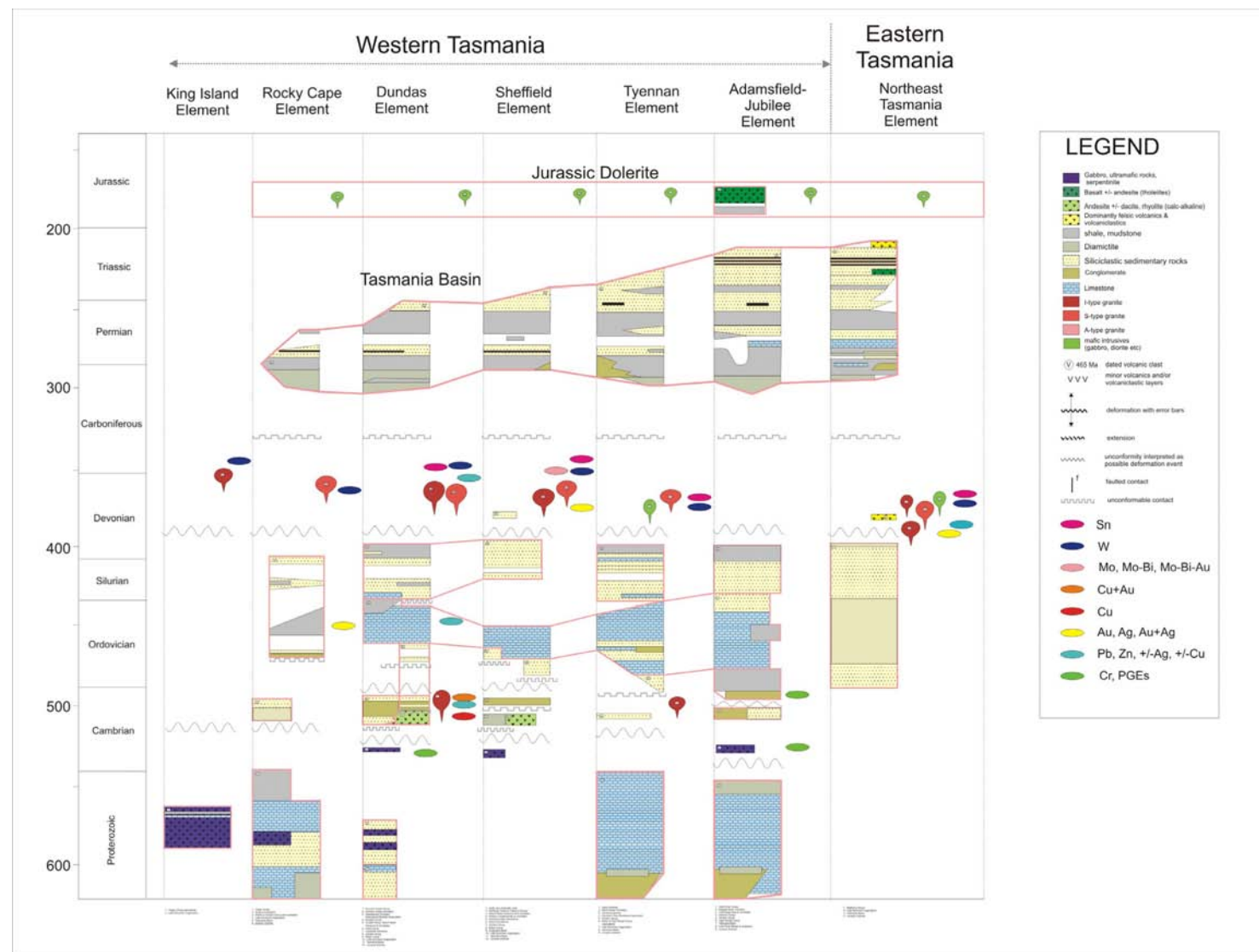


Figure 7. Time-space plot for Tasmania. Modified from, and based largely on, the time-space plot of Seymour and Calver (1998), with modifications by G. Green, J. Everard, C. Calver and M. McClenaghan (Mineral Resources Tasmania, pers. comm., 2008). Refer to Figure 4 for location of elements.

1.1.1: Late Neoproterozoic to Early Cambrian

Rodinia break-up, pre-Delamerian Orogeny (pre-515 Ma)

Delamerian Orogen

Geological and tectonic summary

The Late Neoproterozoic (ca. 600 Ma) to mid Cambrian geological history of southeastern Australia records a cycle of continental rifting and ocean opening, related to the breakup of Rodinia, formation of passive margins and initiation of the Pacific Ocean (e.g., Cawood, 2005). This continued in southeastern Australia until it was effectively ended by subduction (starting at ca. 515 Ma; Foden et al., 2006) and arc-continent collision, ca. 510-505 Ma (Berry and Crawford, 1988; Crawford and Berry, 1992), related to the Delamerian Orogeny. Glen (2005) called this period the Delamerian Cycle, and suggested that it lasted more than 300 Ma, back to ca. 830-780 Ma, and possibly earlier. Most of this time period falls outside the range or geographic coverage of this report and is not covered here (see Drexel and Preiss, 1995; Calver and Walter, 2000; Crawford et al., 2003a; Glen, 2005; and references therein for more information).

Neoproterozoic and Early Cambrian rocks occur in western Tasmania and King Island (Figs 7, 10), and in the Glenelg and Grampians-Stavely Zones in Western Victoria (Figs 6, 10), and are extensively developed in the Delamerian Orogen in South Australia (Fig. 8). They include glacial-derived sediments (ca. 700 and 635 Ma; Calver and Walters, 2000), ca. 780-725 Ma and ca. 600-580 Ma rift-related mafic magmatism (Crawford and Berry, 1992; Holm et al., 2003), and the extensive, largely marine, sedimentation of the Stansbury Basin (Drexel and Preiss, 1996; VandenBerg et al., 2000; Crawford et al., 2003b; Figs 6, 8), which also contains ca. 525 Ma rift-related magmatism. Rocks of this age also include ultramafics and mafic rocks, interpreted as either oceanic floor and/or supra-subduction zone remnants (Crawford et al., 2003a).

Widespread evidence for continental breakup in southeastern Australia, related to Rodinian rifting (e.g., Cawood, 2005; Crawford et al., 2003a), although preserved in rocks ca. 715 Ma, it is also well recorded in rocks ca. 600 Ma in age and younger, in western Tasmania and King Island (e.g., Calver and Walter, 2000; Calver et al., 2004; Meffre et al., 2004), South Australia (e.g., Drexel and Preiss, 1995; Foden et al., 2001), western Victoria (VandenBerg et al., 2000; Crawford et al., 2003b), and the Koonenberry region, western New South Wales (e.g., Crawford et al., 1997; Gilmore et al., 2007). As summarised by Crawford et al. (1997; 2003a, b) many rocks of this age contain alkaline and/or tholeiitic assemblages consistent with rift tectonics and a passive margin and mantle-plume magmatism. Crawford et al. (2003a) suggested the extension was oriented largely northwest-southeast to explain the distribution of rift volcanism at this time.

Geological history and Time-Space plot explanation

- Neoproterozoic mafic magmatism in western Tasmania suggested to reflect rift volcanism (Holm et al., 2003). Ages poorly constrained, probably ca. 725 Ma, but possibly to ca. 780 and older.
- Deformation in western Tasmania - the Wickham deformation – constrained by granites (King Island and northwest Tasmania) to have occurred at ca. ~760 Ma and 777 Ma respectively (Turner et al., 1998).
- Glacially-derived ca. 700 Ma, ca. 640 Ma and ca. 575-582 Ma sediments in northwest Tasmania (Calver and Walters, 2000; Calver et al., 2004; Kendall et al., 2007).
- ca. 600-570 Ma shelf and rift successions (including glacial-derived sediments) in most elements of Western Tasmania (Seymour and Calver, 1995; 1998; Meffre et al., 2004), including the Togari, Success Creek and Weld River Groups (Fig. 7). Many of these, such as the Crimson Creek Formation, contain mafic tholeiitic magmatic rocks including picrites, consistent with rifting and a mantle plume (Crawford and Berry, 1992; Crawford et al., 2003a).
- Deposition of the Cambrian sediments of the Moralan Supergroup in the Glenelg Zone, Victoria (VandenBerg et al., 2000; Crawford et al., 2003b; Figs 6, 10), and more extensively in South

Australia (Drexel and Preiss, 1996; Fig. 8), all as part of the Stansbury Basin. In the Glenelg Zone, the sediments are dominantly deep-water turbidites, but also include mafic magmatic rocks (VandenBerg et al., 2000; Crawford et al., 2003b). Rift-related, within-plate and more oceanic MORB-like magmatism occurs within the Morolana Supergroup, in both South Australia and western Victoria - the Truro Volcanics and correlatives (Drexel and Preiss, 1996, Belperio et al., 1998; VandenBerg et al., 2000; Crawford et al., 2003b; Figs 6, 8). Most of these appear to have ages of ca. 525 Ma. Deposition of the Stansbury Basin appears to have ended by ca. 514 Ma, based on granite emplacement ages (Foden et al., 2006).

- Fault-bounded, mafic-ultramafic complexes in western Victoria (VandenBerg et al., 2000; Crawford et al., 2003b) and in western Tasmania, interpreted as oceanic floor and/or supra-subduction zone remnants (e.g., Crawford and Keays, 1987; Crawford et al., 1984, 2003b; Fig. 7). These mafic-ultramafic complexes either pre-date (VandenBerg et al., 2000), or are contemporaneous with, early Delamerian deformation (e.g., ca. 510 Ma age for the Heazlewood Ultramafic Complex, Tasmania; Turner et al., 1998).

Rodinian break-up, pre-Delamerian Orogeny – Lachlan Orogen

Late Neoproterozoic and Early (to Late) Cambrian rocks of this age also occur within the Lachlan Orogen region (Fig. 9). These fall into two main categories:

- Mafic and ultramafic Cambrian (and older?) igneous rocks, preserved along major faults, especially those that delineate zone boundaries in the Victorian part of the Lachlan (e.g., VandenBerg et al., 2000, Spaggiari et al., 2003, 2004). These are similar to those recorded in the Delamerian Orogen; all have been subdivided into a number of chemical associations, including an ultramafic and a tholeiitic-boninitic association (e.g., Crawford and Keays, 1987; Crawford et al., 1984, 2003b; VandenBerg et al., 2000). These are typically interpreted as having formed in a suprasubduction zone environment (e.g., Crawford and Keays, 1987; Crawford et al., 1984, 2003b; Spaggiari et al., 2003, 2004). In the Stawell Zone, the mafic volcanics (Magdala Volcanics; Figs 9, 10) have a back-arc signature and are underlain by continental-derived turbiditic sediments (Crawford et al., 2003b; Squire et al., 2006). These authors interpreted the succession to represent a distal back-arc environment, related to a west-dipping subduction zone to the east. This is also consistent with the observation that at least some of these mafic-ultramafic successions, e.g., in the Bendigo and Tabberabbera zones, appear to form the basement to those zones (e.g., Spaggiari et al., 2003, 2004; Korsch et al., 2008; Fig. 6), that is, floored by oceanic crust as suggested by Gray, Foster and co-workers (e.g., Gray and Foster, 2004). These are overlain, conformably in places, by Cambrian, deep marine, often pelagic sedimentation (VandenBerg et al., 2000; Spaggiari et al., 2003; Fig. 6). Examples, such as those in the Bendigo Zone and further east, were not affected by the Delamerian Orogeny (Spaggiari et al., 2003, 2004). Western examples including those in the western Stawell Zone were deformed during the Delamerian Orogeny (e.g., Miller et al., 2006).
- Interpreted Delamerian-age and older rocks of the Selwyn Block (VandenBerg et al., 2000; Fig. 9). The Selwyn Block hypothesis suggests the presence of older continental basement beneath the Melbourne Zone, which has been linked to western Tasmania (see Cayley et al., 2002 for detailed discussion). Although some authors have suggested oceanic crust, only, as basement for the Lachlan Orogen (e.g., Gray, 1997; Gray and Foster, 1997, 1998), the presence of the Selwyn Block appears to be confirmed by the recent central Victorian seismic survey (e.g., Korsch et al., 2008; Cayley et al., in prep). The seismic results clearly show the Melbourne Zone to be underlain by something distinct from zones to the west. The Selwyn Block contains Cambrian calcalkaline volcanics (e.g., VandenBerg et al., 2000, Spaggiari et al., 2003; Figs 9, 10), which have many similarities to, and have been correlated with, the Mount Read Volcanics (Crawford et al., 2003a, b). Crawford et al. (2003b) suggested that these volcanics may represent ‘along strike continuations’ of the Mount Read Volcanics. In the oceanic crust basement model, these calcalkaline rocks are interpreted as island arc volcanics, e.g., Jamieson Island Arc (Gray and Foster, 2004; Spaggiari et al., 2003, 2004).

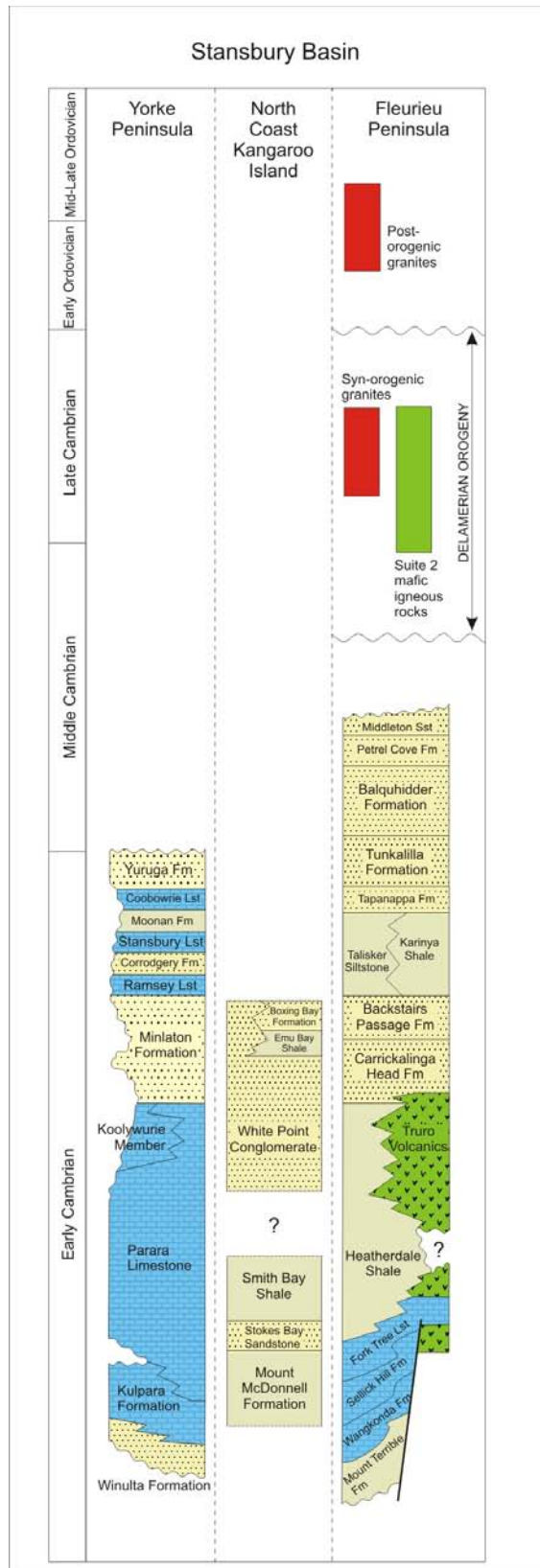


Figure 8. Time-space plot for the Stansbury Basin region and younger granites of South Australia. Modified from Drexel and Preiss (1995).

These associations provide important tectonic constraints. The evidence from rocks of this age in the Bendigo and Tabberabbera Zones and the Selwyn Block shows that parts of the now contiguous Lachlan Orogen were significantly separated in the Early Palaeozoic, as suggested by many authors (e.g., VandenBerg et al., 2000; Cayley et al., 2002; Gray and Foster, 2004). Crawford et al. (2003a) have suggested that the Selwyn Block represented part of Australia rifted off during the 600 Ma event.

1.1.2. Early to Middle Cambrian

Delamerian Orogeny: ca 515 Ma to ca. 490 Ma

Geological and tectonic summary

The break-up of Rodinia and associated extension of southeastern Australia between the Late Neoproterozoic (ca. 600 Ma) and the Early Cambrian was halted with the development of subduction and accompanying contractional orogenesis – the Delamerian Orogeny. In South Australia and western Victoria, the Delamerian Orogeny commenced at ca. 515 Ma (e.g., Foden et al., 2006; Fig. 8). A similar time is evident for the Delamerian Orogeny (sometimes called Tyennan Orogeny) in Tasmania (Seymour and Calver, 1995; Fig. 7).

In South Australia, the Delamerian Orogeny was long-lived, from ca. 515 to 490 Ma (e.g., Drexel and Preiss, 1995; Foden et al., 2006). VandenBerg et al. (2000), amongst others, suggested that there may have been two deformation stages, ca. 515 and ca. 490 Ma, and indicated that, in Victoria, the first phase is evident in the Glenelg Zone, the second only in the Grampians-Stavely Zone (Fig. 6). Miller et al. (2006) showed that the Delamerian Orogeny also affects the western part of the Stawell Zone, based on metamorphic ages of ca. 490–500 Ma. Western Tasmania records at least two discrete deformation events, separated by a significant extensional event. The Delamerian (~Tyennan) Orogeny in western Tasmania was initiated, by collision, around 510 Ma (Berry and Crawford 1988; Crawford et al., 2003a), with subsequent post-collisional (back-arc) extension, and renewed contractional deformation around 495 Ma (Berry, 1994; Turner et al., 1998; Fig. 7).

VandenBerg et al. (2000) showed that, in many respects, western Victoria and western Tasmania share similar geological histories. The Delamerian Orogeny in both western Tasmania and western Victoria was triggered by arc-continent collision around 515–510 Ma (Berry and Crawford, 1992; VandenBerg et al., 2000; Crawford et al., 2003a). In both areas collision was accompanied by the accretion of Cambrian forearc boninitic crust – the Tasmanian mafic-ultramafic complex in Tasmania (Crawford and Berry, 1992; Crawford et al., 2003a), and the Dimboola Igneous Complex in western Victoria (VandenBerg et al., 2000; Crawford et al., 2003b; Figs 6, 7). High temperature, low pressure, metamorphism and metamorphic complexes developed in both regions, with syn-tectonic I- and S-type granites emplaced in the Glenelg River Metamorphic Complex in Western Victoria (VandenBerg et al., 2000; Crawford et al., 2003b). Both regions underwent subsequent post-collisional extension (possibly in a backarc environment), leading to emplacement of calcalkaline volcanics - Mount Read Volcanics, correlatives, and intrusives, in Tasmania (Crawford and Berry, 1992; Crawford et al., 2003a), and the Mount Stavely Volcanic Complex in Victoria (Crawford et al., 1996, 2003b; VandenBerg et al., 2000). This was followed by a second phase of deformation at ca. 500 Ma to 490 Ma. Subsequent extension resulted in rift basins, and deposition of molasse-type sediments, e.g., Owen Conglomerate (Seymour and Calver, 1995) in Tasmania, and emplacement of post-tectonic granites in western Victoria and South Australia (e.g., VandenBerg et al., 2000; Foden et al., 2006; Figs 6, 7, 8).

The polarity of subduction in the Delamerian is uncertain, with both east-dipping (e.g., VandenBerg et al., 2000; Munker and Crawford, 2000; Crawford et al., 2003a) and west-dipping (e.g., Gray and Foster, 2004; Foden et al., 2006) subduction having been invoked. Probably the best evidence for polarity is provided by evidence from the backarc volcanic rocks in the Stawell Zone (see previous section), which Squire et al.

(2006) interpreted to represent a distal backarc environment, related to a west-dipping subduction zone to the east. Similarly, calcalkaline volcanics in western Tasmania and the Melbourne Zone, suggest a west-dipping arc. As noted by Squire et al. (2006), this scenario also explains the observed succession (forearc to backarc) evident in many of the mafic-ultramafic complexes (seafloor remnants) preserved in both the Delamerian and Lachlan Orogens (Crawford et al., 2003b; Fig. 9). The presence of these rocks in the Lachlan Orogen strongly indicates an oceanic setting for much of the Lachlan at this time (e.g., Crawford et al., 1984; Gray, 1997; Fergusson, 2003; Gray and Foster 2004; Spaggiari et al., 2003). Finally, it would appear that additional offshore arcs are also required to explain the arc-related remnants of this age in the New England Orogen (e.g., Aitchison et al., 1994) and the ca. 515 Ma Takaka Island arc in New Zealand (e.g., Munker and Crawford, 2000).

In western Victoria and South Australia, the end of the Delamerian is marked by the end of deformation and emplacement of post-tectonic intrusives ca. 495–485 Ma (Foden et al., 2006; Fig. 8). At this time, subduction is best recorded well to the east, in the Ordovician Macquarie Arc (Crawford et al., 2007a).

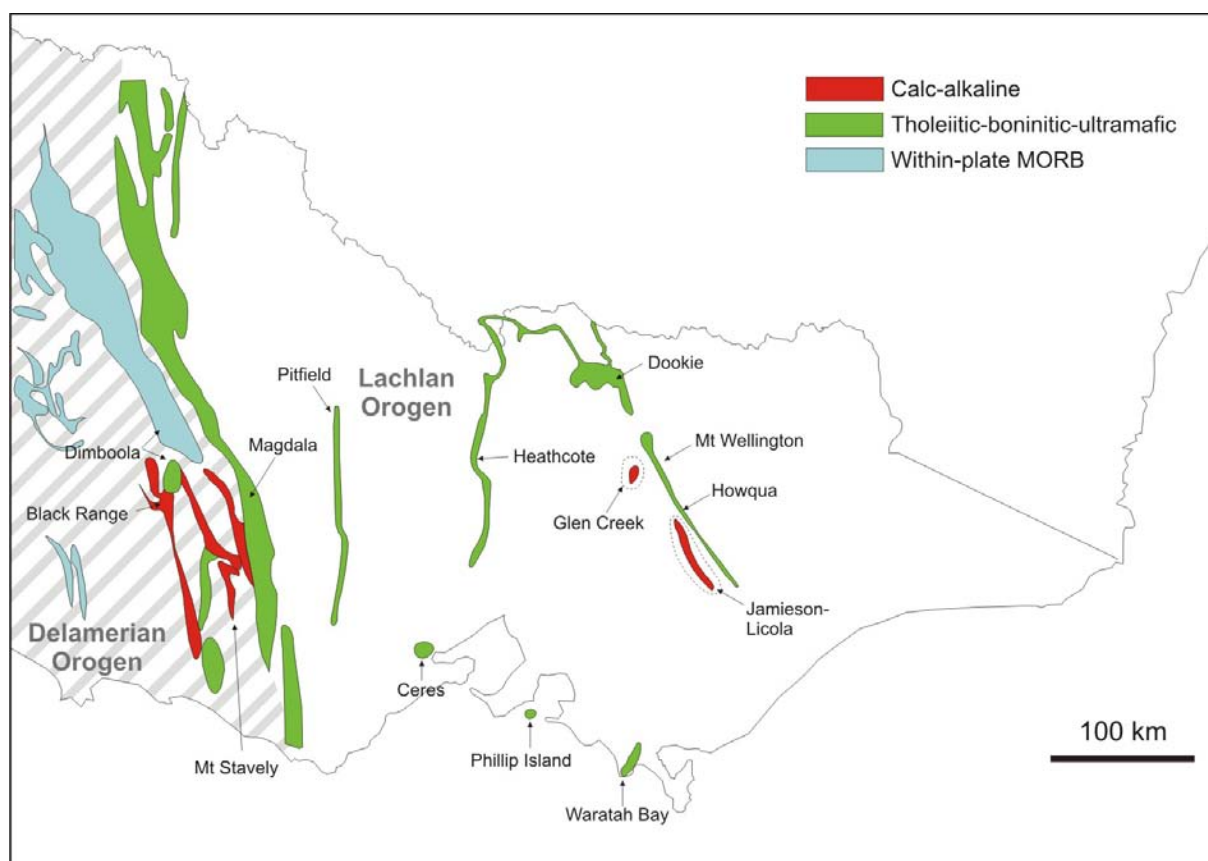


Figure 9. Distribution of various lithological and chemical associations of the Neoproterozoic to Cambrian volcanic rocks of Victoria. Figure modified from VandenBerg et al. (2000). Rocks at Ceres, Phillip Island, Glen Creek and Jamieson-Licola are interpreted to form part of the Selwyn Block (VandenBerg et al., 2000; Cayley et al., 2002). Others (e.g., Gray and Foster, 1997, 2004; Spaggiari et al., 2004) interpret the Jamieson-Licola rocks as remnants of an island arc.

Geological history and Time-Space plot explanation

- Stage 1 Delamerian deformation in Western Victoria (Grampians Zone) and Tyennan deformation in Western Tasmania, related to arc-continent collision in the Early to Middle Cambrian (from ca 520–515 Ma), e.g., Berry and Crawford, 1988; Crawford and Berry, 1992; Turner et al., 1998; VandenBerg et al., 2000 (Figs. 6, 7, 8).

- This collision was accompanied by the emplacement of numerous allochthonous blocks, including mafic (tholeiites, boninites, eclogites), ultramafic and sedimentary rocks (Berry and Crawford, 1988; Crawford and Berry, 1992; Turner et al., 1998; VandenBerg et al., 2000; Fig. 9). These rocks have been considered to have formed, at least partly, in the forearc region of an island arc (Crawford and Berry, 1992). Dating of these rocks indicate Middle Cambrian ages (515-510 Ma; Turner et al., 1998). This is consistent with ages for the obduction event which is bracketed between ca. 515 and ca. 500 Ma (from a tonalite in the ophiolite; Black et al., 1997; and overlying Mount Read Volcanics; Berry and Crawford, 1988; Crawford et al., 2003a). Similar ages are recorded in western Victoria (see VandenBerg et al., 2000; Crawford et al., 2003b).
- Collision was associated with significant deformation and metamorphism, e.g., Ulverstone Metamorphics, Forth Metamorphic Complex, Glenelg River Metamorphic Complex (VandenBerg et al., 2000, Crawford et al., 2003b; Gray et al., 2003; Berry et al., 2007). Berry et al. (2007) recorded ages of ca. 510 Ma for metamorphism in western Tasmania. Syntectonic magmatism occurs in the Glenelg Zone of Victoria, representing the eastern extent of this magmatism best developed in South Australia (e.g., Drexel and Preiss, 1996; Foden et al., 2006).
- Post-collision Middle Cambrian east-west extension (e.g., Crawford and Berry, 1992), possibly in a backarc environment led to formation of the Dundas Trough and the mixed-derivation sedimentary succession of the Lower Dundas Group and correlatives in Tasmania (Seymour and Calver, 1995), and the marine sedimentation of the Nargoon Group in the Grampians-Stavely Zone of Victoria (VandenBerg et al., 2000; Crawford et al., 2003b; Figs 6, 7, 10). Similar aged deep-marine sedimentation occurred within the Stawell Zone at this time (St Arnaud Group; e.g., VandenBerg et al., 2000).
- The mainly felsic, calcalkaline Mount Read Volcanics, and associated volcano-sedimentary successions (Crawford and Berry, 1992; Munker and Crawford, 2000) formed around this time, as did similar rocks (e.g., Mount Stavely Volcanic Complex) in Victoria (VandenBerg et al., 2000; Crawford et al., 1996; 2003b; Squire et al., 2006; Figs 6, 7, 9, 10). In Tasmania, the volcanics were intruded by late dacites and granites. Similar-aged calcalkaline volcanic rocks (e.g., Licola and Jamieson Volcanics) occur as windows in the Melbourne Zone (VandenBerg et al., 2000; Spaggiari et al., 2003; Fig. 9). These have compositions similar to the Mount Read Volcanics of Tasmania (Crawford et al., 1992, 1996, 2003a, b) – in line with the inferred Selwyn Block connection between the Melbourne Zone and western Tasmania (Cayley et al., 2002). In this model, interpreted environments for these rocks are similar, that is, post-collisional, continental-rift setting (Crawford et al., 1996; VandenBerg et al., 2000; Cayley et al., 2002). Gray, Foster and co-workers have invoked an alternate model for the calcalkaline volcanics of the Melbourne Zone. In their ocean floor basalt model, these rocks are interpreted to be remnants of a Cambrian arc system - the Jamieson Island Arc (e.g., Spaggiari et al., 2003, 2004; Gray and Foster, 2004) - structurally incorporated into the Melbourne Zone.
- A second phase of deformation occurred between 505 Ma and 495 Ma in western Victoria, Tasmania and the Selwyn Block (VandenBerg et al., 2000; Seymour and Calver, 1995; Figs 6, 8). In Tasmania, this deformation has been subdivided into an early phase of north-south contraction in the late Middle-early Late Cambrian and east-west contraction in the Late Cambrian (e.g., Berry, 1994; Seymour and Calver, 1995). Berry et al. (2007) recorded no metamorphism of this age in western Tasmania, although they did in offshore samples which they related to the Ross Orogeny in Antarctica.
- Although now reported from the western part of the Stawell Zone (Miller et al., 2006), the Delamerian Orogeny did not affect rocks in the Lachlan Orogen. The deformation is, however, recorded in the Selwyn Block underlying the Melbourne Zone (VandenBerg et al., 2000; Cayley et al., 2002; although see Spaggiari et al., 2003; Fig. 6). The presence of deformed calcalkaline volcanic rocks which have been equated with the Mount Read Volcanics in Tasmania (Crawford et al., 2003a), strongly suggests that deformation was synchronous with the second phase of the deformation as recorded elsewhere. On the basis of the correlation between the Selwyn Block and Tasmania, this deformation has been referred to as the Tyennan Orogeny (VandenBerg et al., 2000; Cayley et al., 2002).

- Further extension (Late Cambrian), possibly commencing between the two Middle-Late Cambrian deformations, resulted in deposition of coarse sediment, e.g., Owen Conglomerate and correlatives, and upper Dundas Group, in Tasmania (Munker and Crawford, 2000; Crawford et al., 2003a; Seymour and Calver, 1995; Figs 7, 10) and post-tectonic magmatism, including A-types, in western Victoria and South Australia, ca. 495–485 Ma (Foden et al., 2006; Figs 6, 8, 10).
- Continuing marine, mostly deep-water turbiditic sedimentation in the Stawell Zone (e.g., St Arnaud Group), and largely pelagic sedimentation in the Bendigo and Tabberabbera zones (e.g., Goldie Chert) of Victoria (VandenBerg et al., 2000; Crawford et al., 2003b; Gray et al., 2003; Figs 6, 10). The Bendigo and Tabberabbera zones contain evidence of deposition in an oceanic environment distant from continental Australia. As summarised by VandenBerg et al. (2000), these include the pelagic sedimentation, the apparently conformable relationship with ocean floor rocks, and the lack of evidence for the Delamerian Orogeny in these zones.

1.1.3. Late Cambrian to earliest Silurian

Post-Delamerian to Benambran Orogeny: ca. 490 Ma to ca. 430 Ma

Geological and tectonic summary

The Lachlan Orogen is dominated by two contrasting rock packages through this time interval:

- Deep water voluminous quartz-rich turbidites of cratonic provenance and pelagic sediments. These occur throughout the Lachlan Orogen, with little or no evidence for volcanic input. They are conformable on oceanic crust in a number of regions.
- Calcalkaline magmatism and volcanoclastics, and marine sediments with common carbonates – largely related to the Macquarie Arc (e.g., Glen, 2004; Crawford et al., 2007a).

These rocks are thought to have been juxtaposed as part of the Benambran Orogeny (e.g., VandenBerg et al., 2000; Glen et al., 2007b), which appears to have occurred in two main pulses, ca. 440 and 430 Ma (e.g., Glen et al., 2007b). Syn-tectonic, largely S-type magmatism accompanied this deformation.

From the late Cambrian to the end of the Ordovician-Early Silurian, most of the Lachlan Orogen was the site of deep marine sedimentation, in NSW, Victoria and Tasmania (Figs 5, 6, 7, 10). These consist of quartz-rich turbiditic successions (Western, Central and Eastern Lachlan) and late Middle to Upper Ordovician black shale-dominated sediments (Central and Eastern Lachlan). In a number of regions, e.g., Bendigo and Tabberabbera zones (e.g., VandenBerg et al., 2000; Fergusson and VandenBerg, 2003; Spaggiari et al., 2003), this sedimentation appears to be conformable upon mafic and ultramafic rocks interpreted as oceanic crust (see previous section). In most areas of the Lachlan Orogen, this deep water sedimentation ended with the Benambran Orogeny. Sedimentation, however, continued in both the Melbourne Zone and northeastern Tasmania, and both regions show no evidence for the Benambran Orogeny (e.g., Fergusson and VandenBerg, 2003; Seymour and Calver, 1995; Figs 6, 7, 10). Most workers imply the presence of one or more submarine fans for the Ordovician deep marine sedimentation (e.g., Fergusson and VandenBerg, 2003; Gray and Foster, 2004; Glen, 2005; Glen et al., 2007b). This may have reflected uplift of the Delamerian Orogen in the early Ordovician (e.g., VandenBerg et al., 2000; Fergusson and VandenBerg, 2003), although there is uncertainty regarding the relative positions of the respective parts of the Lachlan Orogen at this time. The major change in sedimentation recorded by the switch to black shale-dominated pelagic sediments in the late Middle Ordovician and their localisation largely to the Central and Eastern Lachlan is, in part, contemporaneous with early Benambran deformation (ca. 455 Ma) recorded in the Western Lachlan (VandenBerg et al., 2000; Gray et al., 2003; Miller et al., 2006).

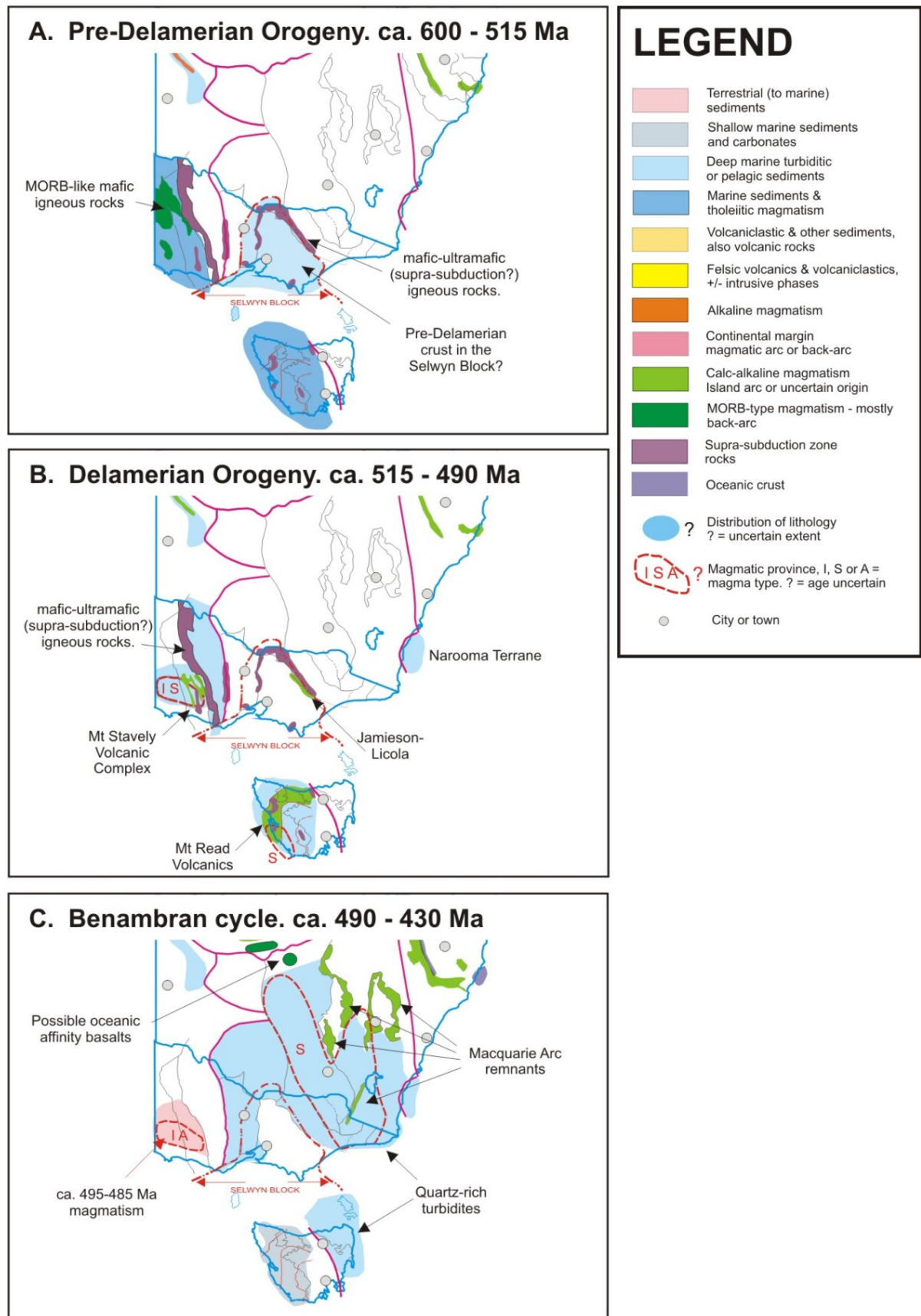


Figure 10. Generalised distribution of rocks in the Lachlan Orogen, by tectonic cycle. A – pre-Delamerian, B – Delamerian Orogeny, C – Benambran cycle.

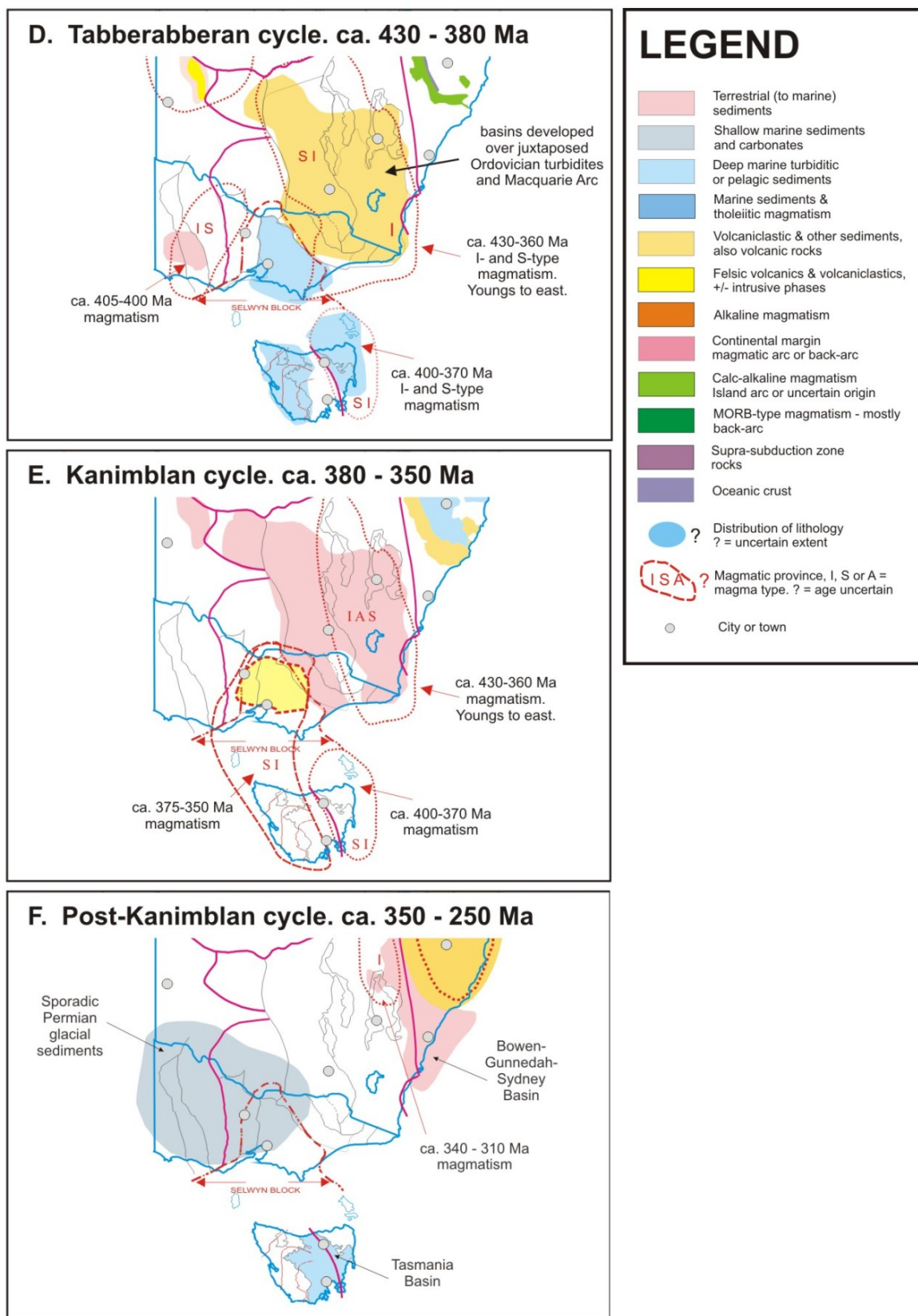


Figure 10 (continued). Generalised distribution of rocks in the Lachlan Orogen, by tectonic cycle. D – Tabberabberan cycle, E – Kanimblan cycle, F – post-Kanimblan to Hunter-Bowen cycle.

These sediments apparently contain no evidence of volcanic detritus, and it would appear, as suggested by numerous workers (e.g., Gray and Foster, 2004; Meffre et al., 2007), that the contemporaneous Macquarie Arc was disconnected from the deep marine sedimentation, perhaps by hundreds of kilometres or more (Meffre et al., 2007). This is consistent with the interpreted oceanic environment for the Macquarie Arc (e.g., Percival and Glen, 2007). The Early Ordovician to earliest Silurian Macquarie Arc consists of calcalkaline and shoshonitic volcanics, intrusions and volcanoclastic and carbonate-rich successions, typically suggested to have formed in an intra-oceanic arc setting (e.g., Crawford et al., 2007a; Figs 2, 5, 10), although Wyborn (1992) suggested the largely shoshonitic magmatism reflected an older, not current, subduction setting. The arc is preserved as four elongate remnants, mostly in New South Wales (Figs 2, 10), though includes the Kiandra Group in northern Victoria (Fergusson and VandenBerg, 2003). Recent detailed work (Crawford et al., 2007a and companion papers) suggests that the Macquarie Arc was built up in four successive phases of growth, with apparently two distinct (east and west) provinces which may not have been together until accretion (Percival and Glen, 2007). The arc is thought to have accreted to eastern Australia in the Early Silurian as part of the Benambran event (Glen et al., 2007b; Meffre et al., 2007).

Contemporaneous shallow marine and terrestrial sedimentation was deposited on top of rocks of the Delamerian Orogen, e.g., in western Tasmania (Seymour and Calver, 1995; Figs 7, 10) and in the Koonenberry region (Gilmore et al., 2007). Post-tectonic A- and I-type magmatism occurred in the Delamerian Orogen, e.g., in the Glenelg Zone, Victoria, and South Australia (Foden et al., 2006). No sedimentation is known from western Victoria at this time.

Deformation associated with the Benambran Orogeny commenced in the Western Lachlan, e.g., Stawell and Bendigo zones, ca. 455-440 Ma (VandenBerg et al., 2000; Gray et al., 2003; Gray and Foster, 2004) and ca. 440 Ma (and slightly older) in other parts of the Lachlan (e.g., Collins and Hobbs, 2001; Gray et al., 2003; Glen et al., 2007b). The resulting Benambran Orogeny affected most of the Lachlan, with the exception of the Melbourne Zone and apparently all of Tasmania (VandenBerg et al., 2000; Seymour and Calver, 1995, 1998). The deformation was accompanied by crustal thickening and uplift, regional, locally significant, metamorphism (in rocks of the Adaminaby Superterrane, such as in the Wagga-Omeo Metamorphic Belt) and it also marked the end of recorded arc volcanism in the Macquarie Arc (Crawford et al., 2007a). Significant higher grade metamorphism (e.g., Gray, 1997; Gray et al., 2003), as well as syn-tectonic S-type magmatism (e.g., Collins and Hobbs, 2001), accompanied the orogeny; both are largely concentrated in two, non-parallel, belts, one in the Central Lachlan, one in the Eastern Lachlan (Fig. 10). The orogeny probably involved at least two discrete deformation events, ca. 440 and 430 Ma (Glen et al., 2007b), and resulted in a complex arrangement of terranes, particularly in eastern NSW. A number of accretion events are inferred to have occurred during the Benambran Orogeny. These include the Macquarie Arc terrane and elements of the Adaminaby Superterrane (Glen, 2005; Glen et al., 2007b) and Benambra Terrane (Willman et al., 2002), as well as the Narooma Terrane (Glen, 2005).

Overall, the Lachlan Orogen appears to record a relatively simple 'passive-margin' to deep marine environment and an interpreted oceanic arc environment. These are often depicted as an oceanic backarc basin (marginal sea) behind an oceanic arc and west-dipping slab (e.g., Coney, 1992; Glen et al., 1998; Cayley et al., 2002; Fergusson, 2003; Gray et al., 2003; Gray and Foster, 2004; Glen, 2005). In detail, however, the situation is more complex and controversy exists the position of terranes, the the number and location of subduction zones and mechanisms of terrane accretion (e.g., Coney, 1992; Gray, 1997; Soesoo et al., 1997 and discussion papers; VandenBerg et al., 2000; Collins and Hobbs, 2001; Cayley et al., 2002; Willman et al., 2002; Cas et al., 2003; Fergusson, 2003; Gray et al., 2003; Spaggiari et al., 2003, 2004; Glen, 2005; Glen et al., 2007b), and even whether subduction was present at all, e.g., Wyborn (1992). Many models invoke a number of subduction zones (e.g., Gray, 1997; Soesoo et al., 1997, Collins and Hobbs, 2001; Fergusson, 2003; Spaggiari et al., 2003, 2004) to explain the across-orogen variations in magmatism, metamorphism, deformation, especially vergence changes, and the presence of blueschists. The actual mechanisms and detail of these reconstructions have important implications for mineralisation, given that the Benambran event coincides with significant lode Au (e.g., Bendigo) and arc-related Cu-Au (e.g., Cadia) mineralisation (e.g., VandenBerg et al., 2000; Crawford et al., 2007a). Tectonic reconstructions and implications for related metallogenesis are discussed in sections 2 and 3.

Geological history and Time-Space plot explanation

- Ordovician deep marine turbiditic sediments occur in the Western, Central and Eastern Lachlan: in Victoria and NSW, e.g., St Arnaud, Castlemaine, Adaminaby, Wagga and Girilambone groups (e.g., VandenBerg et al., 2000; Crawford et al., 2003b; Glen, 2004; Watkins and Meakin, 1996; Lyons et al., 2000; Colquhoun et al., 2005) – called the Adaminaby Superterrane by Glen et al. (2007b); and in Tasmania - the Mathinna Group (Seymour and Calver, 1995; [Figs 2, 5, 6, 7, 10](#)). These are well developed sequences, the major exception being the Melbourne Zone where rocks of this age are poorly developed (Fergusson and VandenBerg, 2003).
- Post-tectonic A- and I-type magmatism, ca. 495-485 Ma, with no accompanying sedimentation, occur within the mainland Delamerian Orogen at this time e.g., in the Glenelg Zone, Victoria, and South Australia (Foden et al., 2006; [Figs 6, 8, 10](#)). Western Tasmania records deep water sedimentation followed by extensive Ordovician carbonate-dominated sedimentation (e.g., Gordon Group) (see below; Seymour and Calver, 1995).
- Late Ordovician, mainly pelagic, black-shale dominated rocks, e.g., Bendoc Group (Lewis et al., 1994; VandenBerg et al., 2000; Seymour and Calver, 1995, 1998; Colquhoun et al., 2005) which appear to be confined to the Central and Eastern Lachlan in NSW and Victoria, and in eastern Tasmania ([Figs 5, 6, 7, 10](#)). Absent from most of the Melbourne Zone, only apparently occurring around the eastern margins (Mount Easton Shale; VandenBerg et al., 2000; Fergusson and VandenBerg, 2003). The switch in sedimentation style, from turbiditic sediments to black shales, appears to be contemporaneous with initial Benambran deformation in the Western Lachlan (e.g., Fergusson and VandenBerg, 2003; Gray et al., 2003). Deep water clastic sedimentation appears at the top of these successions, at least locally (e.g., VandenBerg et al., 2000; Colquhoun et al., 2005).
- Ordovician deep marine sediments overlying mafic volcanics of the Wagonga Group were also deposited during this time in the Narooma Terrane (Lewis et al., 1994; Glen et al., 2004; [Figs 5, 10](#)). The upper part of this succession has an increasing continental component, interpreted to represent the increasing influence of mainland Australia (Glen et al., 2004; Glen, 2005), presumably as the two became closer together. Glen et al. (2004) interpreted the Narooma Terrane as being oceanic in origin; Miller and Gray (1997), in contrast, presented an accretionary wedge model to explain the terrane.
- Ordovician mafic volcanics, chert and serpentinite and other (Cambrian-Ordovician) ultramafic rocks, interpreted as oceanic crust, occur within the Eastern Lachlan in New South Wales, e.g., Jindalee Group (Warren et al., 1995; Lyons et al., 2000; [Fig. 5](#)). As indicated by Glen (2005) and Meffre et al. (2007), these probably represent remnants of the basement to the turbidites, brought up by subsequent deformation, analogous to the older examples in Victoria. Late Ordovician(?) ultramafic rocks also occur in the Western Lachlan, interpreted as intrusive into the Ordovician sediments (e.g., Lyons et al., 2000).
- Shoshonitic and calcalkaline volcanics, intrusions and volcanoclastic and carbonate-rich successions were formed in the Ordovician to earliest Silurian, preserved in four zones: the Junee-Narromine Volcanic Belt, its possible extension the Kiandra Volcanic Belt, the Molong Volcanic Belt, and the Rockley-Gulgong Volcanic Belt ([Figs 2, 5](#)). These rocks have collectively been called the Macquarie Arc (e.g., Glen, 2004; Crawford et al., 2007a). They have been described in detail by Crawford et al. (2007a) and companion papers (special issue of the Australian Journal of Earth Sciences). As outlined by Crawford et al. (2007a, b; Percival and Glen, 2007; Glen et al., 2007a), the Macquarie Arc is believed to represent an intra-oceanic arc setting, which was built up in four successive phases from the Early Ordovician (Phase 1) to the Late Ordovician to earliest Silurian (Phase 4). The arc is thought to have accreted to eastern Australia in the Early Silurian as part of the Benambran event. Percival and Glen (2007) document two distinct (east and west) provinces within the arc, which they suggested were not together until accretion. Subduction zone polarity for the Macquarie Arc is uncertain and may have been either west-dipping, east-dipping or a mixture of east- and west-dipping subduction (e.g., see models in Meffre et al., 2007). These models also include significant strike-slip motion south of the arc. Although general consensus favours an arc interpretation for these rocks, other interpretations have been invoked. Wyborn (1992), for example, suggested the arc-like geochemical signature reflected a previous (Cambrian),

- not current, subduction environment. Wyborn (1992) speculated that a plume-type environment could explain the Ordovician shoshonitic, and subsequent Silurian-Devonian felsic, magmatism.
- Early Benambran Orogeny east-west shortening deformation and east vergence recorded in the Western Lachlan, e.g., Stawell and Bendigo zones, ca. 455-440 Ma (VandenBerg et al., 2000; Gray et al., 2003; Gray and Foster, 2004; Figs 5, 6, 7). This early deformation may be possibly diachronous, commencing earlier in the west, becoming younger to the east (e.g., Gray and Foster, 1997). Major orogeny occurred at ca. 440 Ma (and slightly older) in other parts of the Lachlan (e.g., Collins and Hobbs, 2001; Gray et al., 2003; Glen et al., 2007b), with the exception of Melbourne Zone in Victoria, and the Mathinna Group in Tasmania (Seymour and Calver, 1995). This deformation was accompanied by crustal thickening, uplift, regional, locally significant, metamorphism (in rocks of the Adaminaby Superterrane, such as in the Wagga-Omeo metamorphic belt; Gray, 1997; VandenBerg et al., 2000; Willman et al., 2002; Gray et al., 2003; Gray and Foster, 1997, 2004; Glen et al., 2007b) and also marked the end of recorded arc volcanism in the Macquarie Arc (Crawford et al., 2007a). Significant Au and Cu-Au mineralisation occurred during this event (see Section 3).
 - Late Ordovician to early Silurian turbiditic sedimentation in the Central and Eastern Lachlan in Victoria, e.g., the syn-tectonic Yalmy group (VandenBerg et al., 2000; Willman et al., 2002; VandenBerg, 2003; Glen et al., 2007b; Fig. 6). These rocks have significantly more quartz than underlying rocks and reflect a change in provenance, which VandenBerg et al. (2000) suggested was possibly related to uplift associated with the first phase of the Benambran Orogeny. This was contemporaneous with continuing deep-marine sedimentation in the Melbourne Zone in Victoria, and both western and eastern Tasmania (Seymour and Calver, 1995; see below).
 - Phase 2 of the Benambran Orogeny at ca. 430-425 Ma, best developed in the Eastern Lachlan, expressed as renewed east-west to more oblique contraction to strike-slip deformation (with a north-south component) (e.g., VandenBerg et al., 2000; Collins and Vernon, 2001; VandenBerg, 2003; Pogson and Glen, 2006; Glen et al., 2007b; Figs 5, 6). This is not apparent everywhere, e.g., Colquhoun et al. (2005) indicate deformation in the Cargelligo region of New South Wales had ended prior to ca. 432 Ma post-tectonic S-type granites (although see Collins and Hobbs, 2001). Gray and co-workers (e.g., Gray, 1997; Gray and Foster, 1997; Gray et al., 2003) document a change in vergence from eastwards in the Western Lachlan to south-west vergence in the central Lachlan. These workers invoked a number of subduction zones to explain this structural change (e.g., Gray and Foster, 1997; Soesoo et al., 1997). Willman et al. (2002) suggest that this event was responsible for amalgamation of parts of the Benambra Terrane (Central and Eastern Lachlan), although VandenBerg et al. (2000) indicate younger strike-slip movement (Bindian Orogeny) may be responsible for bringing the central and eastern zones into their final positions.
 - Continuing regional metamorphism, such as in the Wagga-Omeo Metamorphic Belt and Cooma region, accompanied the second phase of the Benambran deformation (Foster and Gray, 1997; Gray et al., 2003), associated with extensive S-type syn- (to post) tectonic magmatism (Collins and Hobbs, 2001). The S-type magmatism is concentrated within two, large, non-parallel belts (e.g., Chappell et al., 1991; Collins and Hobbs, 2001), one in the Central Lachlan, the other in the Eastern Lachlan (Fig. 10). Collins and Hobbs (2001) suggested that at least some of the thermal input was provided by mantle-derived magmatism. They suggested, in a variant of the Gray and Foster (1997) model, that two magmatic arcs were required to explain these belts. Both models invoke an east-northeast dipping subduction under the Tabberabbera Zone (central Lachlan).
 - The end result of the Benambran Orogeny was effectively cratonisation of much of the Western Lachlan (VandenBerg, 2003), and complex accretion of a number of terranes in the eastern part of the Orogen (VandenBerg et al., 2000; Willman et al., 2002; Glen, 2005; Glen et al., 2004, 2007b; Meffre et al., 2007). Neither phase of deformation appears to have significantly affected the Melbourne Zone in Victoria (VandenBerg et al., 2000), and the Mathinna Group in Tasmania (Seymour and Calver, 1995) – possibly reflecting structural partitioning. Both the latter regions were the sites of continued marine sedimentation through this period (Figs 6, 7).
 - Similarly, sedimentation in Western Tasmania continued through this time. Sedimentation in western Tasmania began with post-Delamerian extension, resulting in formation of rift or sag basins, with deep water sediments including conglomerates (e.g., Owen Conglomerate) ca. 490 to

470 Ma (e.g., Seymour and Calver, 1995). These were followed by extensive Ordovician carbonate-dominated sedimentation (e.g., Gordon Group) and a switch to deeper water clastic-dominated sedimentation (e.g., Eldon Group) around the end of the Ordovician (Seymour and Calver, 1995, 1998; Figs 7, 10). Deposition continued intermittently until the Middle Devonian, and the Benambran and Bindian orogenies appear to be absent. Hiatuses recorded in several elements (e.g., Sheffield element; Fig. 7), however, may correspond to the Benambran Orogeny (Seymour and Calver, 1995, 1998), as may the switch from carbonate- to clastic-dominated sedimentation. It is possible that the apparent absence of these orogenies may simply reflect the presence of the Selwyn Block, such that deformation was largely partitioned around it (e.g., VandenBerg et al., 2000). Reed (2001) has inferred an earlier Benambran deformation in the older western part of the Mathinna Supergroup. Detrital zircon data has been interpreted to suggest western and eastern Tasmania were separate during this period (e.g., Black et al., 2004).

1.1.4. Middle Silurian to Middle to early Late Devonian

Post-Benambran to Tabberabberan Orogeny: ca. 430-380 Ma

Geological and tectonic summary

This time period in the Lachlan Orogen (Figs 5, 6, 7, 10) is marked by widespread extensional episodes with accompanying basin formation and ubiquitous extrusive and intrusive magmatism, possibly related to significant arc roll-back after the Benambran Orogeny (e.g., Glen et al., 2004; Spaggiari et al., 2004). Differing geological histories are observed within the Orogen, in particular, between the Western Lachlan (Whitelaw Terrane) and Central and Eastern Lachlan (Benambra Terrane), suggesting these regions were separate for most of this cycle (e.g., Gray and Foster, 1997, 2004; VandenBerg et al., 2000; Willman et al., 2002; Spaggiari et al., 2004). In addition, the regions of the Selwyn Block (Cayley et al., 2002) – the Melbourne Zone in Victoria, and western Tasmania - also record unique aspects in their geology.

Within the Central and Eastern Lachlan, extension developed within and across the juxtaposed Ordovician turbidite successions and the Macquarie Arc remnants in New South Wales (Meakin and Morgan, 1999; Lyons et al., 2000; Glen et al., 2007b; Figs 5, 10), and within Ordovician turbidite successions in Victoria (e.g., VandenBerg et al., 2000; Figs 6, 10). This resulted in widespread largely deep to shallow marine sedimentation (including carbonates) in troughs, platforms/shelves and other rift zones (Pogson and Watkins, 1998; Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; VandenBerg, 2003; Colquhoun et al., 2005; Glen, 2005; Figs 5, 6). Accompanying magmatism includes S- and I-type granites, and is often bimodal – intermediate magmatism is uncommon (Chappell and White, 1992; Lyons et al., 2000; Collins and Hobbs, 2001; Gray et al., 2003; Rossiter, 2003; Black et al., 2005).

These basins were inverted during the poorly-defined, latest Silurian (-Earliest Devonian) Bindian Orogeny (ca. 420-410 Ma; Figs 5, 6). This orogeny, most evident in Victoria and New South Wales, has been suggested to be transpressive and to have resulted in significant strike-slip movement between the western and central Lachlan (VandenBerg et al., 2000; Willman et al., 2002; Glen, 2005), with possibly up to 600 km of dextral movement (e.g., Willman et al., 2002). Other authors have questioned this, and it is possible that deformation of this age may relate more to continuing subduction-accretion effects if the multiple subduction zone models of Gray (1997), Gray and Foster (1997), Soesoo et al. (1997), Spaggiari et al. (2003, 2004), are correct. This does not necessarily negate strike-slip effects in the Central and Eastern Lachlan. Bindian deformation appears to relate more to east-west contraction in far eastern Victoria (e.g., Willman et al., 2002). Renewed extension and development of rift basins continued in the Early Devonian in the Central and Eastern Lachlan following the Bindian Orogeny (e.g., Willman et al., 2002). This resulted in deep to shallow marine sedimentation (including carbonates) and widespread bimodal and felsic volcanism in new and existing basins and rifts in the Central and Eastern Lachlan in both Victoria and New South Wales (Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000, Willman et al., 2002; Colquhoun et al., 2005; Glen, 2005).

The Bindian Orogeny appears to be absent from northeastern Tasmania and the Melbourne Zone of Victoria (Seymour and Calver, 1995; VandenBerg et al., 2000; Figs 6, 7). As a result, dominantly deep marine sedimentation in both regions is largely continuous throughout this cycle, continuing from Benambran times (e.g., VandenBerg et al., 2000; Fergusson et al., 2003; Seymour and Calver, 1995). In the Melbourne Zone, sedimentation appears to both shallow upwards to terrestrial sedimentation, as well as contain evidence for a change in sediment transport direction and source, with the appearance of lithic and volcanoclastic detritus derived from the east (VandenBerg et al., 2000; VandenBerg, 2003). Willman et al. (2002) and VandenBerg (2003) suggested these changes are evidence for the arrival of the Benambra Terrane, that is, the Benambran and Whitelaw terranes were separated prior to this – a conclusion agreed upon by many workers, regardless of tectonic model (e.g., Gray, 1997; Gray and Foster, 1997, 2004; Fergusson, 2003; Spaggiari et al., 2004). Similarly, detrital zircon data from sediments of this age in northeastern Tasmania indicate no apparent sourcing of material from western Tasmania, which led Black et al. (2004) to suggest that northeastern and western Tasmania were also separate at this time.

Rocks of this age also occur in the Delamerian Orogen. These include terrestrial to marine sedimentary rocks in western Victoria, largely in the Grampians-Stavely Zone (VandenBerg et al., 2000), and deep water clastic-dominated sedimentation in Western Tasmania (Seymour and Calver, 1995, 1998; Figs 6, 7, 10). Sediments in Western Victoria were apparently deformed ca. 420–410 Ma (Bindian Orogeny?) and are overlain by post-deformation Early Devonian volcanic rocks (VandenBerg et al., 2000) and associated plutonism. The Bindian Orogeny appears to be absent in western Tasmania, although hiatuses in sedimentation which may correspond to this orogeny are recorded (Seymour and Calver, 1995, 1998).

Widespread felsic-dominated magmatism occurs across the Western, Central and Eastern Lachlan within this cycle (Lyons et al., 2000; Collins and Hobbs, 2001; Chappell and White, 1992; Rossiter, 2003; Black et al., 2005; Gray et al., 2003; Figs 5, 6, 7, 10). Crystallisation ages largely fall between ca. 430 and 390 Ma but include younger granites belonging to the Kanimblan cycle (e.g., Chappell and White, 1992; Gray et al., 2003; Black et al., 2005). The oldest granites in New South Wales and Victoria are dominantly S-types in the Central and Eastern Lachlan (e.g., Collins and Hobbs, 2001; Willman et al., 2002). These include a continuation of the magmatism that commenced during the Benambran Orogeny (e.g., Collins and Hobbs, 2001). Early Silurian magmatism appears to be absent from the Western Lachlan (Whitelaw Terrane) in Victoria (VandenBerg et al., 2000; Willman et al., 2002), although some may occur in the Glenelg Zone (VandenBerg et al., 2000). Granite ages in the Lachlan orogen appear to correlate with geography. In the Eastern Lachlan of New South Wales and Victoria, ages appear to decrease eastwards, to ca. 380 to 360 Ma (Lewis et al., 1994; VandenBerg et al., 2000), most probably reflecting arc roll-back. In Victoria, however, granites appear to largely decrease in age towards the Melbourne Zone. Granites east and west of the Melbourne Zone are largely ca. 420 to 380 Ma in age (VandenBerg et al., 2000; Gray et al., 2003; Rossiter, 2003; Figs 6, 10). The youngest rocks occur within the post-Tabberabberan Middle Devonian to Early Carboniferous (ca. 385–350 Ma) Central Victorian Magmatic Province (VandenBerg et al., 2000; Rossiter, 2003). A similar diachronous trend is evident in Tasmania, where granite ages record a pronounced westward younging in crystallisation age from ca. 400–375 Ma, pre-, syn-, and post-tectonic granites in the northeast, to post-tectonic granites, ca. 370–350 Ma in western Tasmania (Black et al., 2005; Figs 7, 10). The large areal distribution of magmatism in the Lachlan Orogen, and the thermal requirements, are problematical and led many authors to speculate on possible tectonic scenarios. Suggested tectonic environments include multiple subduction zones (e.g., Collins and Hobbs, 2001; Soesoo et al., 1997; Gray and Foster, 1997), delamination (Collins and Vernon, 1994), mantle plumes (e.g., Wyborn, 1992; Cas et al., 2003), as well as backarc extension. Both the mantle plume and the multiple subduction models, e.g., the double subduction model of Soesoo et al. (1997), have the advantage of explaining the wide distribution of the Lachlan Orogen granites. The multiple subduction zone models, unlike the plume models, can explain the diachronous trends of the granites, but are, however, largely not consistent with the bimodal or dominantly felsic nature of the magmatism (however, cf. Collins and Hobbs, 2001). Cas et al. (2003) suggested magmatism effectively ceased (went into hiatus) just before the Tabberabberan Orogeny. Similarly, it is difficult to see how the plume model could explain the geographic spread and trends in ages of magmatism.

The ca. 390-380 Ma east-west contractional Tabberabberan Orogeny (e.g., Gray and Foster, 1997, 2004; Spaggiari et al., 2003, 2004), effectively cratonised the whole Lachlan Orogen. The orogeny has suggested to have been responsible for the final amalgamation of the terranes and zones of the Lachlan, e.g., Whitelaw (Western Lachlan) and Benambran (Central and Eastern Lachlan) terranes (Gray and Foster, 1997, Soesoo et al., 1997; VandenBerg et al., 2000; Willman et al., 2002; Spaggiari et al., 2003, 2004; Figs 5, 6), and western and eastern Tasmania (Seymour and Calver, 1995; Black et al., 2005; Fig. 7). Interpretations for the drivers of this east-west contraction are varied, but largely reflect the different interpreted tectonic models for the region. Gray and co-workers (e.g., Gray, 1997; Gray and Foster, 1997, 2004; Soesoo et al., 1997; Spaggiari et al., 2003, 2004) suggested that this collisional event was (at least partly) related to the closure of a marginal basin (effectively the Melbourne Zone) and an end to double divergent subduction (e.g., Gray and Foster, 1997; Soesoo et al., 1997). Conversely, Willman et al. (2002) and Cayley et al. (2002) suggested the Tabberabberan Orogeny was responsible for ending the relative strike-slip movement of the Whitelaw and Benambran Terranes, and reflected docking of the two terranes. In Tasmania, the Tabberabberan deformation (ca. 388 Ma; Black et al., 2005) may relate to docking of northeastern Tasmania to western Tasmania (Black et al., 2004). This deformation may also relate to docking of the Calliope-Gamilaroi arc in the New England Orogen at this time. In reality, it is possible that all of these models may be partly correct, as each model better explains some, but not all, aspects of the Tabberabberan geology. Despite this apparent complexity, it would appear that, as for the Benambran cycle, the Tabberabberan cycle records a relatively simple overall backarc environment behind a west-dipping slab and subduction zone located to the east (e.g., Gray, 1997; Glen et al., 1998; Cayley et al., 2002; Gray and Foster, 2004; Glen, 2005; Collins and Richards, 2008).

Geological history and Time-Space plot explanation

- Deposition of widespread, largely deep marine to shallow marine, sediments (including carbonates) in dominantly north-south oriented troughs, platforms, shelves and other rift zones (e.g., Cowra, Tumut, Rast, Jemalong, Hill End troughs, Cobar Basin, Walter Range, Mumbil Shelves, Cowombat Rift, e.g., Pogson and Watkins, 1998; Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; VandenBerg, 2003; Colquhoun et al., 2005; Glen, 2005), across the Lachlan Orogen in NSW and eastern parts of Victoria (Figs 5, 6, 10). This was accompanied by widespread, largely S-type, but also I-type and possibly A-type (Colquhoun et al., 2005), volcanism and plutonism. This is thought to reflect post-Benambran east-west extension or transtension, and formation of rift basins (e.g., VandenBerg et al., 2000; Willman et al., 2002; Colquhoun et al., 2005; Glen, 2005), suggested to be related to easterly roll-back migration of the arc following the Benambran Orogeny (e.g., Gray and Foster, 2004; Glen et al., 2004; Spaggiari et al., 2004). Extension developed within and between the Ordovician turbidites and the Macquarie Arc (Lyons et al., 2000; Glen et al., 2007b). Glen (2005) suggested that volcanism is absent from basins north of the Lachlan Transverse Zone, which he suggested reflected poorly understood differences in the thermal structure of the region.
- In Victoria, most deposition of this age was in the Melbourne Zone, continuing from pre-Benambran times (VandenBerg et al., 2000; Fergusson et al., 2003; Figs 6, 10). This consists of (latest Ordovician-) Silurian to Middle Devonian dominantly deep marine turbiditic sedimentation of the Murrindindi Supergroup (VandenBerg et al., 2000; VandenBerg, 2003), which is dominated by mudstone-siltstone but is marked by episodic influxes of coarser clastic turbidites. Sedimentation appears to shallow upwards (VandenBerg et al., 2000), with carbonates occurring towards the top and, locally preserved, very thick terrestrial sedimentation (Cathedral Group) at the very top of the Supergroup (Fig. 6). VandenBerg (2003) suggested that the Cathedral Group was deposited possibly in response to the start of the Tabberabberan Orogeny. VandenBerg (2003) also indicated a change in sediment transport direction and source in the Emsian with the appearance of lithic and volcanoclastic detritus derived from the east. This has been taken as evidence for the incoming arrival of the Benambran Terrane, possibly by strike-slip motion, to something approaching its present position (VandenBerg et al., 2000; Willman et al., 2002). Prior to this, an open ocean appears to have existed east of the Melbourne Zone (Whitelaw Terrane).
- In western Victoria, largely in the Grampians-Staveland Zone, deposition of terrestrial to marine sediments of the Grampians Group (VandenBerg et al., 2000; VandenBerg, 2003; Figs 6, 10)

occurred on top of rocks of the Delamerian Orogen. Part of this sedimentation appears to be contemporaneous with the Yalmy Group in eastern Victoria, that is, it is partly syn-Benambran in age (VandenBerg et al., 2000; VandenBerg, 2003). The Grampians Group was apparently deformed ca. 420-410 Ma and is overlain by post-deformation Early Devonian volcanic rocks (VandenBerg et al., 2000; VandenBerg, 2003) and associated plutonism. This deformation has been equated with the Benambran Orogeny (VandenBerg et al., 2000) but appears to be very similar in age to the Bindian Orogeny in eastern Victoria. Miller et al. (2006) document a low strain southeast-northwest contraction occurring at this time (ca. 420-414 Ma) in both the Grampians Group and in the Stawell Zone.

- Marine turbiditic sedimentation in the Mathinna Group in northeastern Tasmania, continued from the previous cycle, probably up to Early Devonian, although age constraints for this succession are poor (Seymour and Calver, 1995; Figs 7, 10). The group was deformed in the Early Devonian (Reed, 2001). Detrital zircon data from the Mathinna Group indicates no sourcing of material from western Tasmania (Black et al., 2004). These authors suggested this indicated that northeastern and western Tasmania were separate at this time.
- Following the switch to deeper water clastic-dominated sedimentation (e.g., Eldon Group) around the end of the Ordovician (Seymour and Calver, 1995, 1998), deposition in Western Tasmania (Figs 7, 10) continued intermittently until the Middle Devonian. Both the Benambran and Bindian orogenies appear to be absent within this terrane. Hiatuses are recorded in several zones (e.g., Sheffield Element; Fig. 7), however, which may correspond to one or both of these orogenies (Seymour and Calver, 1995, 1998).
- Widespread felsic-dominated magmatism, including the majority of granites in the orogen, occurred across the Western, Central and Eastern Lachlan within this cycle (Meakin and Morgan, 1999; Lyons et al., 2000; Collins and Hobbs, 2001; Chappell and White, 1992; Gray et al., 2003; Rossiter, 2003; Black et al., 2005; Figs 5, 6, 7, 10). Ages largely fall between ca. 430 and 390 Ma, though magmatism continued after the Tabberabberan Orogeny into the Kanimblan cycle, until the early Carboniferous (ca. 360-350 Ma), e.g., Chappell and White (1992), Gray et al. (2003), Black et al. (2005). The oldest granites of the Tabberabberan cycle in New South Wales and Victoria are dominantly S-types in the Central and Eastern Lachlan (e.g., Collins and Hobbs, 2001; Willman et al., 2002). Early Silurian magmatism of this age appears to be absent from the Western Lachlan (Whitelaw Terrane) in Victoria (VandenBerg et al., 2000; Willman et al., 2002; Gray et al., 2003), although some may occur in the Glenelg Zone (VandenBerg et al., 2000). Ages for intrusive magmatism within the Lachlan Orogen show a variety of geographically-controlled diachronous trends. In the Eastern Lachlan of New South Wales and Victoria (Kuark and Mallacoota Zones), ages appear to decrease eastwards (Figs 5, 6, 10), with the youngest granites - Late Devonian in age (ca. 380 to 360 Ma; Lewis et al., 1994; VandenBerg et al., 2000; Gray et al., 2003) - occurring in the Bega and Moruya batholiths (Bega basement terrane of Chappell et al., 1988). In Victoria, granites appear to mostly decrease in age towards the Melbourne Zone (VandenBerg et al., 2000; Gray et al., 2003; Rossiter, 2003). Granites east and west of the Central Victorian Magmatic Province are largely ca. 420 to 380 Ma in age (e.g., Gray et al., 2003). Willman et al. (2002) suggested that these granites are post-tectonic in the Whitelaw Terrane, but, in part, syn-tectonic (related to the Bindian Orogeny) in the Benambra Terrane. The youngest rocks occur within the Middle Devonian to Early Carboniferous (ca. 385-350 Ma) Central Victorian Magmatic Province (Chappell et al., 1988; VandenBerg et al., 2000; Rossiter, 2003; Fig. 6), corresponding to the Melbourne and Bassian Terranes of Chappell et al. (1988). A similar diachronous trend is evident in Tasmania, where granites ages record a pronounced westward younging in granite age from ca. 400-375 Ma, pre-, syn-, and post-tectonic, granites in the northeast to post-tectonic granites, ca. 370-350 Ma in western Tasmania (Black et al., 2005; Figs 7, 10). The large areal distribution of magmatism in the Lachlan Orogen is problematical, and has led many authors to speculate on tectonic scenarios, including multiple subduction zones (e.g., Collins and Hobbs, 2001; Soesoo et al., 1997; Gray and Foster, 1997), delamination (Collins and Vernon, 1994), and mantle plumes (e.g., Wyborn, 1992; Cas et al., 2003). Cas et al. (2003) suggested magmatic activity effectively ceased (went into hiatus) just before the Tabberabberan Orogeny.

- Deformation during the Bindian Orogeny appears to be largely confined to the southern parts of the Central and Eastern Lachlan (e.g., Gray, 1997; Gray and Foster, 1997; Willman et al., 2002; Gray et al., 2003; Glen, 2005), although Miller et al. (2006) record contractional deformation of this age in the Stawell Zone. In eastern Victoria, ages of ca. 418-410 Ma (e.g., Gray and Foster, 1997; Gray et al., 2003) are largely expressed as angular unconformities, especially above the early rift basins of the Tabberabberan cycle (VandenBerg et al., 2000; VandenBerg, 2003). According to VandenBerg et al. (2000), there are no significant effects in western Victoria. Miller et al. (2006), however, document deformation of this age (ca. 420-414 Ma) in the Stawell Zone, which they correlated with similar aged deformation in the Grampians Group in western Victoria. VandenBerg et al. (2000), however, related this deformation to late effects of the Benambran Orogeny. In New South Wales, the Bindian Orogeny, where observed, appears to be of similar age to that recorded in Victoria, e.g., Glen (2005) summarised evidence for ca. 418 to 410 Ma deformation. The deformation, however, is not recorded everywhere in New South Wales; for example, apparently continuous sedimentation during this period is recorded in the Hill End, Cowra and Rast Troughs and Walter Range Shelf (Pogson and Watkins, 1988; Meakin and Morgan, 1999; Colquhoun et al., 2005; Fig. 5). Elsewhere in New South Wales, the deformation, although not recorded, may correspond to breaks in sedimentation, e.g., Early and Late Silurian sedimentation in the Central Lachlan (e.g., Lyons et al., 2000); notably, however, no significant unconformities are recorded in these areas (Lyons et al., 2000). Importantly, there are also unconformities of this age, e.g., in the Kandos and Queens Pinch Groups, which have been suggested to relate to local effects such as uplift associated with granite intrusion (Meakin and Morgan, 1999). The Bindian Orogeny is also not recorded in Tasmania (e.g., Black et al., 2005). The Bindian deformation has been suggested to be transpressive and relate to significant strike-slip movement between the Western and Central Lachlan (VandenBerg et al., 2000; Willman et al., 2002; Glen, 2005). Willman et al. (2002), for example, suggested up to 600 km of dextral movement between the Whitelaw and Benambra Terranes, largely accommodated along what they called the Baragwanath Transform, with most effects within the western part of the Benambra Terrane. Most of this strike-slip movement is thought to have occurred during the Bindian Deformation (Willman et al., 2002). Other workers (e.g., Spaggiari et al., 2003), find no evidence for significant strike-slip movement, at least along the Whitelaw-Benambra Terrane boundary. In addition, it is evident that deformation of this age in the Western Lachlan (Victoria) may relate more to continuing subduction-accretion effects if the multiple subduction zone models of Gray (1997), Gray and Foster (1997), Soesoo et al. (1997) and, more recently, by Fergusson (2003) and Spaggiari et al. (2003, 2004), are correct. This does not necessarily negate strike-slip effects in the Central and Eastern Lachlan, although deformation appears to relate more to east-west contraction in far eastern Victoria (Kuark and Mallacoota zones; Willman et al., 2002).
- Renewed extension and rifting, in the Early Devonian, with deposition of deep marine to shallow marine, sedimentation (including carbonates) in new and existing basins and rifts in the Central and Eastern Lachlan in both Victoria and New South Wales, e.g., Hill End Trough, Buchan Rift (Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; Willman et al., 2002; Colquhoun et al., 2005; Glen, 2005; Figs 5, 6, 10). This was accompanied by widespread bimodal and felsic volcanism, e.g., Gregra Group, Snowy River Volcanic Group (Pogson and Watkins, 1998; Meakin and Morgan, 1999; VandenBerg et al., 2000; VandenBerg, 2003; Cas et al., 2003), as well as associated intrusions. This Early Devonian sedimentation and volcanism is thought to reflect post-Bindian orogenic extension, and appears to be reflected even in basins that do not record the Bindian Orogeny (e.g., Pogson and Watkins, 1998; Meakin and Morgan, 1999; Glen, 2004). Watkins (in Meakin and Morgan, 1999) suggested that the volcanism reflected a backarc environment.
- During the Late Early Devonian to Middle Devonian (ca. 400-380 Ma), the Lachlan Orogen underwent approximately east-west contractional deformation and associated, mostly low-grade metamorphism, related to the Tabberabberan Orogeny (Seymour and Calver, 1995, 1998; Gray, 1997; Gray and Foster, 1997, 2004; Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; Willman et al., 2002; Gray et al., 2003; Spaggiari et al., 2003; Black et al., 2005; Glen, 2005; Figs 5, 6, 7). This deformation, the first to affect most of the Lachlan Orogen, was apparently responsible for the amalgamation of many of the provinces of the Lachlan, and resulted

in much of the current configuration, that is, it effectively cratonised the orogen. In Victoria, the Tabberabberan deformation was responsible for the deformation and uplift of the Melbourne Zone and the amalgamation of the Whitelaw and Benambra terranes (Gray, 1997; VandenBerg et al., 2000; Willman et al., 2002; Spaggiari et al., 2003). Spaggiari et al. (2003) suggested that this collisional event occurred ca. 400-390 Ma, and was related to the closure of the marginal basin (effectively the Melbourne Zone) and an end to double divergent subduction (e.g., Gray and Foster, 1997; Soesoo et al., 1997). Conversely, Willman et al. (2002) suggested the Tabberabberan Orogeny was responsible for ending the relative strike-slip movement of the Whitelaw and Benambra terranes, and reflected docking of the two terranes (Cayley et al., 2002). In Tasmania, the Tabberabberan deformation may relate to docking of northeastern and western Tasmania, based on detrital zircon evidence (Black et al., 2004). Black et al. (2005) record an age of ca. 388 Ma for this event. Reed (2001) documented two apparent phases of Tabberabberan deformation in northeast Tasmania, with opposite sense of vergence. Reed (2001) suggested that northeastern Tasmania better correlated with the Tabberabbera Zone of Victoria. Finally, it is evident that the Tabberabberan Orogeny may also, at least in part, relate to the docking of the Calliope-Gamilaroi island arc in the New England Orogen, which is thought to have occurred around this time (Flood and Aitchison, 1992; Murray et al., 2003).

1.1.5. Late Middle Devonian to Late Devonian-Early Carboniferous

Post- Tabberabberan Orogeny to Kanimblan Orogeny: ca. 380-350 Ma

Geological and tectonic summary

The post-Tabberabberan time period in the Lachlan Orogen is marked by widespread extension, rifting and accompanying basin formation, as well as significant extrusive and intrusive magmatism, occurring within a now largely cratonic land mass (VandenBerg et al., 2000; Lyons et al., 2000; Glen, 2005; [Fig. 10](#)). As outlined by Willman et al. (2002), for example, the post-Tabberabberan sedimentary and volcanic rocks overlie major faults and interpreted suture zones belonging to the Tabberabberan Orogeny, with little evidence for significant later reactivation. Like the earlier orogenic cycles, the post-Tabberabberan extension is thought to reflect behind-arc processes related to renewed roll-back, with the main subduction zone and arc now found within the New England Orogen (e.g., Glen, 2005; Collins and Richards, 2008).

Rocks of this age include Middle to Late Devonian sediments accompanied by A-type and bimodal extrusive and intrusive magmatism (e.g., Meakin and Morgan, 1999; Lyons et al., 2000; Lewis et al., 1994; Wormald et al., 2004; [Figs 5, 6, 10](#)), probably related to initiation of post-Tabberabberan extension and associated rifting. This was followed by Lachlan-wide (New South Wales and Victoria, but not Tasmania) late Middle to Late Devonian to Early Carboniferous, clastic, mostly continental, sedimentation, including red beds, of the 'Lambie facies', reflecting continuing, more widespread extension (Lewis et al., 1994; Warren et al., 1995; Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; Cas et al., 2003; Glen, 2004; [Figs 5, 6, 10](#)). Middle Devonian to earliest Carboniferous intrusive I-, S- and A-type magmatism, ca. 380 to 350 Ma (Chappell and White, 1992; Gray et al., 2003; Wormald et al., 2004; Black et al., 2005), occurred throughout this cycle, across the Lachlan Orogen, including Tasmania ([Figs 5, 6, 7, 10](#)).

Extension was terminated by the early Carboniferous (ca. 360-340 Ma) contractional east-west Kanimblan Orogeny (e.g., Gray, 1997; Meakin and Morgan, 1999; VandenBerg et al., 2000; Gray et al., 2003; Glen, 2005). This orogeny folded and inverted Kanimblan cycle and older rocks. It occurred across the Lachlan Orogen, into the Delamerian Orogen (e.g., Gilmore et al., 2007), but is best expressed in the Eastern Lachlan (Gray, 1997; Gray et al., 2003; Glen, 2005).

There is little evidence for arc-related magmatism during this period in the Lachlan, with most evidence clearly indicating that the arc was located further east in the New England Orogen (e.g., Meakin and Morgan, 1999; Glen, 2005; Collins and Richard, 2008).

Geological history and Time-Space plot explanation

- Middle Devonian sedimentation and locally voluminous, bimodal or felsic I- and/or A-type magmatism within the Eastern Lachlan in New South Wales, e.g., Dulladerry Rift, Rocky Ponds Group, Boyd Volcanic Complex, Mount Wellington Volcanic Group (Lewis et al., 1994; Warren et al., 1995; Pogson and Watkins, 1998; Meakin and Morgan, 1999; Lyons et al., 2000; [Fig. 5](#)). Volcanism and sedimentation is thought to be related to post-Tabberabberan rifting (e.g., Lewis et al., 1994; Meakin and Morgan, 1999; Lyons et al., 2000).
- Late Middle to Late Devonian - Early Carboniferous clastic, shallow-marine to mostly continental, sedimentation, including red beds, of the 'Lambie facies' occur across the Lachlan Orogen in Victoria and New South Wales, e.g., Mulga Downs, Harvey, Catombal, Merimbula Groups, Avon Supergroup, Combyingar Formation (Lewis et al., 1994; Warren et al., 1995; Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; Cas et al., 2003; Glen, 2004; Wormald et al., 2004; [Figs 5, 6, 10](#)). These are thought to be related to widespread post-Tabberabberan extension (e.g., Lewis et al., 1994; Meakin and Morgan, 1999; Lyons et al., 2000). In places the sediments are accompanied by, locally voluminous, bimodal or felsic I- and/or A-type magmatism, e.g., Mount Wellington Volcanic Group (VandenBerg et al., 2000; Cas et al., 2003; Wormald et al., 2004; [Figs 5, 6](#)). The sediments are mostly quartz-rich (e.g., Meakin and Morgan, 1999; VandenBerg et al., 2000; Glen, 2005), and Glen (2005) suggested they may have been sourced largely from central Australia, related to uplift associated with the Alice Springs Orogeny. VandenBerg et al. (2000) suggest sediments of this age in eastern Victoria were probably locally derived from 'Tabberabberan highlands'. Angular unconformities and folding occur throughout this succession in Victoria (e.g., VandenBerg et al., 2000). These may relate to local deformation or perhaps extension. Sedimentary rocks of this age are rare in Tasmania (Seymour and Calver, 1995, 1998).
- Post-tectonic felsic-dominated intrusive and extrusive magmatism occurred across parts of the Western, Central and Eastern Lachlan during this cycle (Chappell and White, 1992; Rossiter, 2003; Black et al., 2005; Gray et al., 2003; [Figs 5, 6, 7, 10](#)). Granite types include I-, S- and, relatively common, A-types (Collins et al., 1982; Chappell et al., 1988; VandenBerg et al., 2000; Rossiter, 2003; Wormald et al., 2004). Ages largely fall between ca. 380 to 360 Ma, but magmatism continued locally until the end of the Devonian-Early Carboniferous (ca. 360-350 Ma), e.g., Chappell and White (1992), Gray et al. (2003), Wormald et al. (2004), Black et al. (2005). Granite ages show a variety of diachronous trends. In the Eastern Lachlan of New South Wales, ages appear to decrease eastwards, with the youngest granites - Late Devonian in age (ca. 380 to 360 Ma; Lewis et al., 1994) - occurring in the Bega and Moruya Batholiths. In Victoria, granites appear to mostly decrease in age towards the Melbourne Zone, from both the east and west sides (VandenBerg et al., 2000; Willman et al., 2002; Gray et al., 2003; Rossiter, 2003). The youngest rocks occur within the Middle Devonian to Early Carboniferous (ca. 385-350 Ma) Central Victorian Magmatic Province (VandenBerg et al., 2000; Rossiter, 2003), corresponding to the Melbourne and Bassian Terranes of Chappell et al. (1988). A similar diachronous trend is evident in Tasmania, where granite ages record a pronounced westward younging from ca. 400-375 Ma, pre-, syn-, and post-tectonic granites in the northeast to ca. 370-350 Ma post-tectonic granites in western Tasmania (Black et al., 2005). Causes of this diachronous behavior are not well understood (see discussion for the Tabberabberan cycle).
- Latest Devonian to Early Carboniferous east-west shortening and associated low-grade regional metamorphism of the Kanimblan Orogeny in New South Wales and Victoria (e.g., Gray et al., 2003; Glen, 2004; [Figs 5, 6](#)). This is best expressed in the Eastern Lachlan, with folding and inversion of the Middle to Late Devonian basins (Pogson and Watkins, 1998; Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; Gray et al., 2003; Glen, 2005), although it extends west to the Delamerian Orogen (e.g., Gilmore et al., 2007). Glen (2005) suggests an age of ca. 340 Ma for this deformation in New South Wales. In Victoria the timing is poorly constrained, but must be post-earliest Carboniferous (e.g., VandenBerg et al., 2000). Gray (1997) and Gray and Foster (1997) suggests ages of ca. 360-340 Ma for the Kanimblan Orogeny.

1.1.6. Middle Carboniferous to Latest Permian

Post-Kanimblan to Hunter-Bowen Orogeny: ca. 350 Ma to 250 Ma

Geological and tectonic summary

The Kanimblan Orogeny was the terminal event in the Lachlan Orogen (e.g., Pogson and Watkins, 1998), and subsequent geology appears to relate more to the New England Orogen to the east. From the latest Carboniferous to Permian, Eastern Australia was dominated by tectonic extension and rifting, and many intracratonic basins were initiated in this time, as was the backarc Sydney-Gunnedah-Bowen basin system (see Thomson Orogen section). Within Victoria and New South Wales, the only significant magmatic event was the Late early Carboniferous to Late Carboniferous (ca. 340-310 Ma) intrusive magmatism recorded as a north-northwest belt in the northeastern Lachlan (e.g., Pogson and Watkins, 1998; Meakin and Morgan, 1999; [Figs 5, 10](#)). This consists of I-type, mostly felsic granites of the Bathurst Batholith and the Gulgong Suite (Bathurst basement terrane of Chappell et al., 1988). They are of similar age and geochemistry to volcanic and intrusive rocks in the New England Orogen (e.g., Chappell et al., 1988; Pogson and Watkins, 1998), and may relate to continental arc formation, although Meakin and Morgan (1999) suggest emplacement in an extensional environment, presumably behind the continental arc of the New England Orogen. The eastern part of the Lachlan Orogen is overlain by the latest Carboniferous to Triassic Sydney-Gunnedah-Bowen basin system, which initially developed as a backarc rift behind the New England Orogen (e.g., Korsch et al., in press a; see New England Orogen section). The basin rocks overlie the Carboniferous granites of the Bathurst-Gulgong area. Of similar age to the Sydney-Gunnedah-Bowen Basin are the Parmeener Supergroup sediments of the Tasmania Basin in Tasmania ([Figs 7, 10](#)), which developed over the suture (Tamar Fracture) between western and northeastern Tasmania (Seymour and Calver, 1995, 1998). These consist of a lower succession of glacial and marine sediments and an upper succession (Late Permian and younger) of non-marine sediments including coal measures (Seymour and Calver, 1995, 1998). Remnants of possibly more widespread Permian glacial and marine sedimentation are also recorded (outcrop and sub-surface, e.g., beneath the Murray Basin) in Victoria and southern New South Wales (O'Brien et al., 2003; [Fig. 10](#)).

1.2. North Queensland region, and eastern parts of the Mesoproterozoic Georgetown basement

by DC Champion

Introduction

The North Queensland Orogen (terminology of Glen, 2005), and constituent provinces, sub-provinces and geographic regions are as defined in Figures 11, 12; the time-space plot is shown in Figure 13. The Charters Towers and Barnard regions, which may be part of either the North Queensland or the Thomson orogens are included within the former. The Georgetown and Coen regions, which represent the Proterozoic Australian cratonic margin are also here included in the North Queensland Orogen. This is considered necessary as a considerable part of the Palaeozoic tectonic regime (Tasman Orogen) and subsequent development was focused across this old margin.

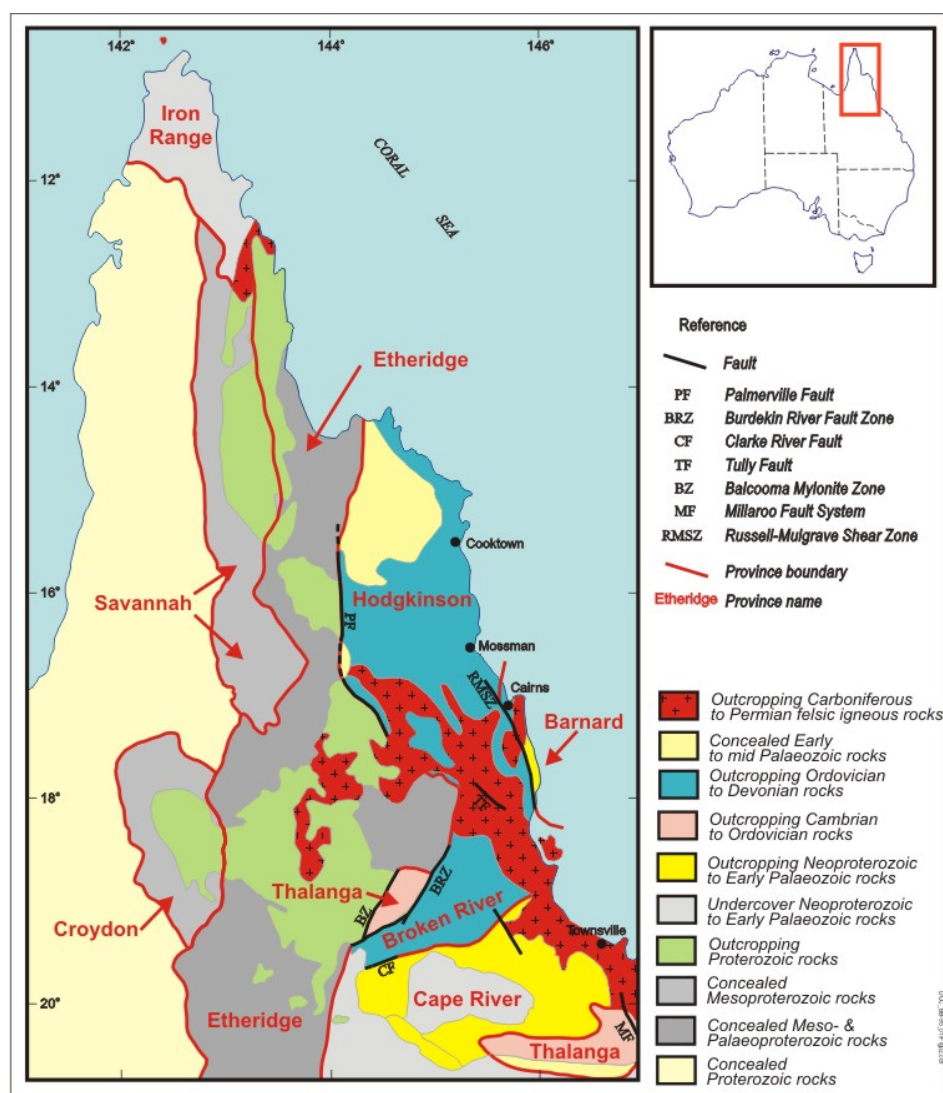


Figure 11. Distribution of geological provinces in northern Queensland. Province names and boundaries as defined by Bain and Draper (1997). The magmatic Macrossan (Cambrian-Ordovician), Pama (Silurian-Devonian) and Kennedy (Carboniferous-Permian) Provinces are not shown. In the Time-Space plots for northern Queensland, Provinces are grouped into regions (see Fig. 12).

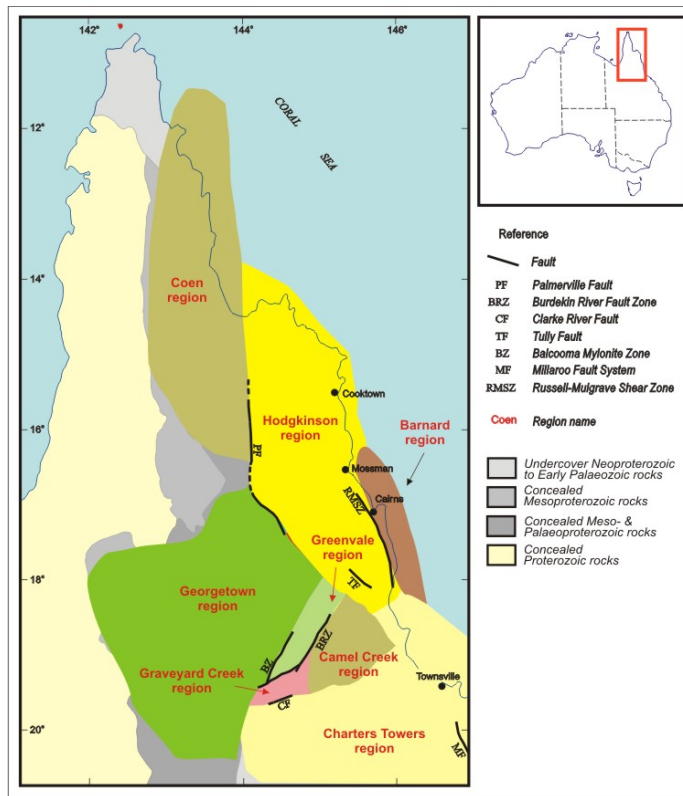


Figure 12. Distribution of geological regions as used for the North Queensland Time-Space Plot (Fig. 13). Geological regions are largely based on the geographic subdivisions used in Bain and Draper (1997). Regions have been used here for areas which include more than one geological province. This approach was largely required because of the large and widespread magmatic Macrossan (Cambrian-Ordovician magmatism), Pama (Silurian-Devonian magmatism) and Kennedy (Carboniferous-Permian magmatism) Provinces which occur across north Queensland and within all of the sedimentary/metamorphic provinces shown in Figure 11.

1.2.1 Late Neoproterozoic to early Cambrian

Rodinian break-up (pre-Delamerian): ca. 600 Ma to 515 Ma

Geological and tectonic summary

Rocks of this age in north Queensland are best represented in the Greenvale Subprovince and Charters Towers region, but also occur in the Georgetown and Coen regions and in the Barnard Province (Figs 13, 14). Most sedimentary successions have poor age constraints. The tectonic environment for this period is typically interpreted as a passive margin, related to (post-dating) Rodinian break-up. Fergusson et al. (2007a, b) suggested the presence of Late Neoproterozoic rifting – ca. 600 Ma – in the region, based on the presence of common 600-500 Ma detrital zircons and on correlation with other regions, e.g., in the Thomson Orogen. The recent results of Fergusson et al. (2007a) also indicate that the Tasman Line, as previously defined, along the eastern margin of the Greenvale Subprovince, does not represent Rodinia breakup, but rather younger (Benambran?) west-directed thrusting of younger rocks over Mesoproterozoic basement.

Geological history and Time-Space plot explanation

Greenvale region

- Metasedimentary-dominated Halls Reward Metamorphics (Withnall et al., 1997a). Age uncertain, but has Cambrian metamorphic ages (ca. 520-510 Ma) and Neoproterozoic detrital zircon ages (Nishiya et al., 2003; Figs 13, 14).

- Mafic to ultramafic magmatic rocks of the Boiler Gully and Gray Creek Complexes, intimately associated with the Halls Reward Metamorphics. Uncertain origin – may be tectonically emplaced and/or intrusive. Withnall et al. (1997a) suggested it was intrusive, at least in part. Ages are poorly constrained and could even postdate the Delamerian Orogeny.
- Late Neoproterozoic-Early Palaeozoic sedimentation (dolomitic carbonate and quartzofeldspathic sediments) of the Oasis Metamorphics (Figs 13, 14). Most probably shallow-marine environment formed as part of the passive Gondwanan margin (Withnall et al., 1997a; Fergusson et al., 2007a). The Oasis Metamorphics include tholeiitic mafic igneous rocks (intrusive/extrusive; e.g., Withnall et al., 1997a). Age constrained to between 540-520 Ma (youngest detrital zircon population) and ca. 485 Ma (overprinting metamorphism; Fergusson et al., 2007a).

Georgetown region

- Terrestrial sedimentary rocks of the Inorunie Group (Withnall et al., 1997a; Figs 13, 14). Ages are poorly constrained, and the group may be as old as Mesoproterozoic – the preferred age of Withnall et al. (1997a).

Barnard region

- Barnard Metamorphics (metasedimentary rocks, amphibolite, metavolcanics) and the intrusive(?) Babalangee Amphibolite (Bultitude et al., 1997; Figs 13, 14). Ages are poorly constrained but must be older than Early Ordovician (ca. 490 Ma), the age of cross-cutting granites (Bultitude et al., 1997).

Coen region

- Sefton Metamorphics – poorly age-constrained metasedimentary rocks (Blewett et al., 1997; Figs 13, 14) and mafic magmatic rocks. Detrital zircons indicate the succession is younger than ca. 1200 Ma (Blewett et al., 1997).

Charters Towers region

- Widespread remnants of largely marine metasedimentary rocks (Charters Towers Metamorphics, Argentine Metamorphics, Running River Metamorphics, Cape River metamorphics; Hutton et al., 1997; Fig. 14). Also includes mafic magmatic rocks, which appear to include both alkaline and tholeiitic compositions (Hutton et al., 1997; Fergusson et al., 2007b). Ages are not well constrained except for the Argentine Metamorphics which Fergusson et al. (2007b) dated as ca. 500 Ma (certainly older than 480 Ma). The Cape River Metamorphics give similar minimum ages, based on intrusive contacts (> ca. 490 Ma; Hutton et al., 1997).

1.2.2. Early to Middle Cambrian

Delamerian Orogeny: ca. 515 to 490 Ma

Geological and tectonic summary

The Delamerian Orogeny is poorly represented in north Queensland, partly reflecting the geochronological uncertainty of many of the units. The best evidence for this orogeny appears to be in the Greenvale Subprovince (metamorphic ages of ca. 520-510 Ma; Nishiya et al., 2003; Fig. 14) and in the Charters Towers region (metamorphics ages of ca. 495 Ma; Fergusson et al., 2007a). Potential Delamerian deformation events may occur in the Georgetown and Coen regions but geochronological control is missing. A number of deformations that could be interpreted as Delamerian have been shown recently, at least partly, to represent post-Delamerian extension (Fergusson et al., 2007a, b).

Geological history and Time-Space plot explanation

Greenvale region

- The Halls Reward Metamorphics have Cambrian metamorphic ages of ca. 520-510 Ma (Nishiya et al., 2003; Fig. 13), consistent with Delamerian deformation. Fergusson et al. (2007a) did not, however, record evidence for Delamerian deformation in the nearby, pre-Delamerian, Oasis Metamorphics.
- Rocks of the Boiler Gully and Gray Creek Complexes appear to have similar deformation histories as the Halls Reward Metamorphics, and so the deformation observed in these complexes is inferred to be Delamerian in age.

Georgetown region

- Poorly age-constrained, widespread, north-south contractional(?), deformation and associated metamorphism (largely retrogressive) in the eastern Georgetown region (Withnall et al., 1997a; Fergusson et al., 2007a; Fig. 13). According to Withnall et al. (1997a) deformation is constrained to between ca. 970 Ma and early Palaeozoic.

Barnard region

- The Barnard Metamorphics record a deformation and metamorphic event (locally high-grade) not observed in the Early Ordovician granites that intrude the metamorphics (Garrahd and Bultitude, 1999; Fig. 13). This is the only time constraint.

Coen region

- Poorly age-constrained deformation and associated metamorphic (greenschist or lower grade) events (Fig. 13). According to Blewett et al. (1997) this deformation is constrained to between ca. 1550 Ma and Devonian (granite emplacement) for most of the Coen Region, but between ca. 1130 Ma and Carboniferous for the Iron Range region (Sefton Metamorphics). The major fabric in the Sefton Metamorphics is east-west (Blewett et al., 1997). Metamorphism is strongest near the Devonian granites (R. Blewett, pers. comm., 2008), suggesting metamorphism may be this age.

Charters Towers region

- Fergusson et al. (2007b) document a ca. 495 Ma contractional deformation event with associated metamorphism in the Argentine Metamorphics (largely overprinted by slightly younger extension). Hutton et al. (1997) and Fergusson et al. (2007b) document similar deformation in the Charters Towers, Running River, and Cape River Metamorphics.

1.2.3. Middle Cambrian to Ordovician-earliest Silurian

Post-Delamerian to Benambran Orogeny: ca. 490 Ma to ca. 430 Ma

Geological and tectonic summary

This time period in North Queensland is dominated by three general supracrustal successions, and a major magmatic province:

- Lower to Middle Ordovician volcanic- or volcanoclastic-dominated assemblages with a calcalkaline signature, interpreted as backarc successions, e.g., Seventy Mile Range Group, Balcooma Metavolcanic Group, part of the Lucky Creek Group (Figs 13, 14).
- Deep water (turbiditic), dominantly, quartz-rich sediments, locally with tholeiitic magmatic rocks, e.g., the Lucky Creek Group (Figs 13, 14).
- Calcalkaline volcanic and carbonate-dominated successions interpreted, in part, as continental or island arc successions.

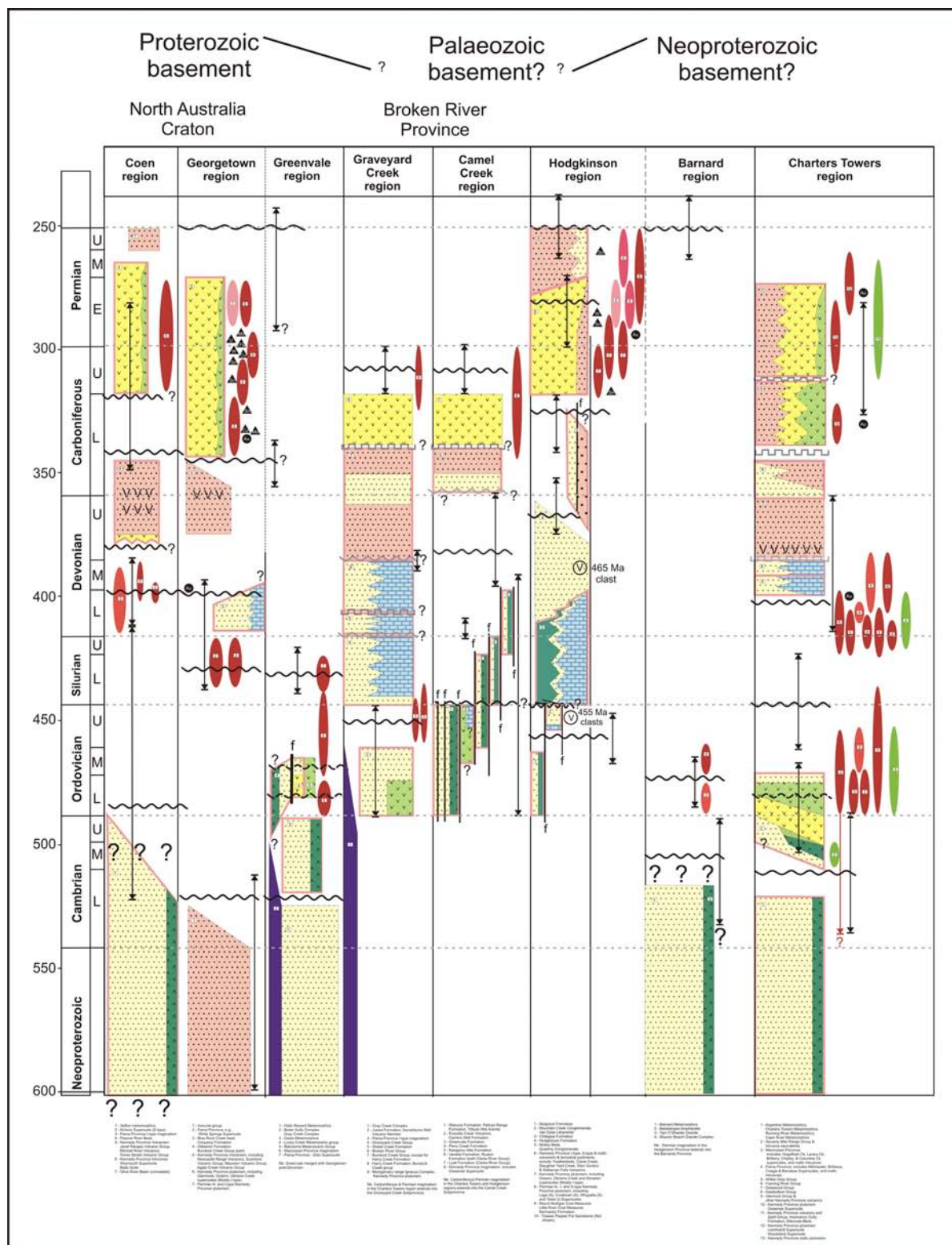


Figure 13. Late Neoproterozoic to Permian time-space plot for the north Queensland region, covering the North Queensland Orogen and Proterozoic basement to the west. Refer to text for data sources and discussion. Regions are as outlined in Figs 11 and 12. Legend over page.

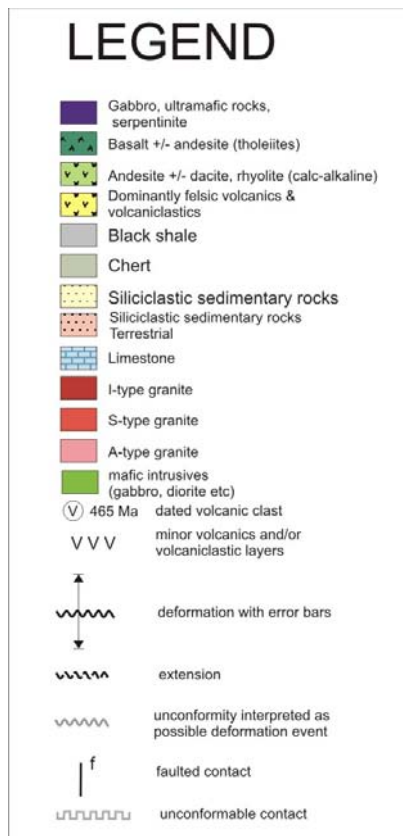


Figure 13. Late Neoproterozoic to Permian time-space plot for the north Queensland region - continued. Legend.

- Mafic to felsic magmatic rocks – the Macrossan Province of Bain and Draper (1997), widely distributed throughout north Queensland but best represented in the Charters Towers region. Magmatic ages range from ca. 490 Ma to ca. 455 Ma (e.g., Hutton et al., 1997; Figs 13, 14). Dominated by I-type and mantle-derived magmatism, but some S-types have been recorded.

Rocks of this cycle have long been interpreted as a dismembered continental margin (e.g., Henderson, 1987). Many authors have suggested backarc, continental or island-arc affinities (e.g., Withnall et al., 1991, 1997b; Henderson, 1986; Stolz, 1994), suggesting an environment not dissimilar to Lachlan Orogen rocks of the same age (e.g., Gray and Foster, 2004; Glen, 2005). Like the Lachlan Orogen, north Queensland rocks of this age are either quartz-rich sediments or calcalkaline volcanics (Fig. 14). Notably, however, unlike the Lachlan Orogen, there are north Queensland units, such as the Judea Formation (Withnall and Lang, 1993), which contain both quartz-rich marine sediments and calc-alkaline volcanics, suggesting proximity between arc-related volcanism and craton-derived sedimentation.

Many of the north Queensland rocks were deformed in the Early Silurian (Fig. 13) by a shortening event coupled with metamorphism – referred to as the Benambran Orogeny by Fergusson et al. (2007a). Evidence for this deformation is found in the Georgetown region where it is constrained by ca. 430 Ma magmatic ages for syn-deformational I-type magmatism (Withnall et al., 1997a; Fergusson et al., 2007a). A similar deformation is recorded in the Charters Towers region, possibly ca. 440 Ma (Fergusson et al., 2007b). Deformation of this age appears to be largely absent in both the Hodgkinson and Broken River Provinces, although Fergusson et al. (2007a) suggested that island-arc terranes within the Camel Creek Subprovince (Broken River Province) - the Everetts Creek Volcanics and Carriers Well Formation – were accreted at this time. Contractional deformation is present within the Hodgkinson and Broken River

provinces but appears to be earlier – probably Late Ordovician (Fig. 13). Garrad and Bultitude (1999) have suggested there may be a time break between Ordovician and Silurian rocks in the Hodgkinson Province that corresponds to uplift (Benambran Orogeny) in the Georgetown region to the west.

Fergusson et al. (2007a, b) documented extensional deformation coupled with low-P high-T metamorphism, greenschist to amphibolite-facies, in both (interpreted) backarc successions in the Greenvale and Charters Towers regions (Fig. 13). They dated these events at ca. 475 Ma and 480 Ma, but suggested extension was in operation from ca. 490 to 460 Ma. This metamorphism and extension was synchronous with granite emplacement, and most of the Macrossan Province magmatism is of this age. Importantly, the interpreted backarc volcanic rocks in the Charters Towers region and Greenvale Subprovince have quite different orientations – ~east-west versus north-northeast–south-southwest, respectively (e.g., Bain and Draper, 1997). This clearly suggests either some later relative movement between the regions, either due to deformation (e.g., Bell, 1980; Fergusson et al., 2007a) and/or perhaps the volcanism formed independently on different crustal fragments. Regardless, given a backarc origin for the volcanic rocks in the southern Charters Towers region, it is possible that Macrossan Province magmatism in the northern part of that region represents the actual magmatic arc (e.g., Henderson, 1980).

Geological history and Time-Space plot explanation

Greenvale region

- Metasedimentary to metavolcanic Lucky Creek Group and Balcooma Metavolcanic Group. Both groups include calcalkaline volcanics though the Lucky Creek Group also contains tholeiitic rocks (Withnall et al., 1997a; Figs 13, 14). The Balcooma Metavolcanic Group is dated at ca. 470 to 480 Ma (e.g., Withnall et al., 1991). Age is uncertain for the Lucky Creek Group, but thought to be the same age, at least in part, as the Balcooma Metavolcanic Group. Withnall et al. (1991) originally suggested the calcalkaline volcanics were part of a volcanic arc, while Fergusson et al. (2007a) suggested a backarc environment. The latter is probably more consistent with similar interpretations for the Charters Towers area (Draper and Bain, 1997).
- Ordovician extension and related metamorphism and granite magmatism dated at ca. 485 to 477 Ma (Fergusson et al., 2007a; Fig. 13).
- Benambran age east-west shortening and metamorphism, and associated syn-deformation I-type plutonism (Withnall et al., 1997a; Fergusson et al., 2007a); granite magmatism has been dated at ca. 430 Ma (Withnall et al., 1997a; Fig. 13).

Georgetown region

- Benambran east-west shortening and metamorphism, and syn-tectonic I-type granite magmatism dated at ca. 430 Ma (Withnall et al., 1997a; Fig. 14). A second deformation event, tentatively dated at ca. 400 Ma (see next section), may also form part of the Benambran event (Withnall et al., 2007a).

Barnard region

- Macrossan Province S- and I-type magmatism, dated at ca. 485 and 464 Ma (Garrad and Bultitude, 1999; Fig. 14).
- Regional deformation and associated low to, possibly high-grade metamorphism with poor age constraints (Bultitude et al., 1997). The deformation event affects the local granites, implying a maximum age of ca. 460 Ma (Garrad and Bultitude, 1999).

Coen region

- Appears to record no definitive evidence of a Silurian deformation event. A poorly age-constrained deformation and associated metamorphic (greenschist or lower) event, constrained to between ca. 1130 Ma and Devonian (Blewett et al., 1997) may be of this age. Like the Georgetown region, deformation in the Coen region is associated with Pama Province magmatism. However, unlike

Georgetown, this deformation and magmatism is largely younger – ca. 407 Ma (post-Benambran) - based on magmatic ages (Blewett et al., 1997).

Charters Towers region

- The latest Cambrian(?) to largely Early Ordovician Seventy Mile Range Group, consisting of a lower succession of marine metasedimentary rocks (with little volcanic input) and an upper assemblage dominated by calcalkaline mafic to felsic volcanics and volcanoclastics (e.g., Henderson, 1986; Hutton et al., 1997; Figs 13, 14). The group is commonly interpreted as forming in a backarc environment (e.g., Henderson, 1986; Stolz, 1994).
- Deformation and associated metamorphism related to both extension (post-Delamerian) and younger north-south Benambran contractional deformation (Berry et al., 1992; Hutton et al., 1997; Fergusson et al., 2005, 2007; Fig. 13).
- Magmatic rocks of the Macrossan Province (Fig. 14). In the Charters Towers region these include granites of the widespread I-type Hogsflesh Creek and Lavery Creek Supersuites, as well as mafic intrusives. Hutton et al. (1997) suggested that the intrusives fall into 2 ages – Late Cambrian to Early Ordovician, and mid Ordovician. The mafic intrusives are, in general, poorly constrained and may be as young as Devonian. The majority of granites post-date the Benambran deformation (Fergusson et al., 2005).

Hodgkinson region

- Marine, quartz-rich, turbiditic sediments with metabasalt of the Mulgrave Formation, preserved in fault-bounded blocks along the Palmerville Fault (Bultitude et al., 1997; Garrad and Bultitude, 1999; Figs 13, 14). Thought to be Early Ordovician in age.
- Limestone, and quartzofeldspathic deep water sediments, of Late Ordovician age, preserved in fault-bounded blocks along the Palmerville Fault (Bultitude et al., 1997; Garrad and Bultitude, 1999). Dacitic volcanic clasts, in conglomerate (Mountain Creek Conglomerate), have been dated at ca. 455 Ma (Garrad and Bultitude, 1999). Garrad and Bultitude (1999) suggested correlation of these units with the limestone and volcanics of the Carriers Well Formation and Everett Creek Volcanics (Broken River region).
- Middle(?) Ordovician east-west shortening. This is only evident in the Early Ordovician(?) Mulgrave Formation (Garrad and Bultitude, 1999), and so is constrained to be older than Late Ordovician (Fig. 13).

Graveyard Creek and Camel Creek regions

- Marine, quartz-rich, turbiditic sediments, commonly with tholeiitic metabasalt, in both the Graveyard Creek and Camel Creek Subprovinces (Withnall and Lang, 1993; Withnall et al., 1997b; Fig. 14). Ages are generally poorly constrained but thought to be Ordovician (Withnall and Lang, 1993; Withnall et al., 1997b), though may extend into the Silurian. The Judea Formation appears to be of Early Ordovician age (Withnall and Lang, 1993).
- Calcalkaline volcanics, volcanoclastics and limestone in both the Graveyard Creek and Camel Creek Subprovinces (Arnold and Fawcner, 1980; Withnall and Lang, 1993; Withnall et al., 1997b; Fig. 14). Ages are variable; Early Ordovician in the Graveyard Creek Subprovince, and Late Ordovician in the Camel Creek Subprovince. The calcalkaline rocks of the latter subprovince were suggested by Fergusson et al. (2007a) to represent island-arc remnants. A number of units appear to contain both quartz-rich marine sediments and calcalkaline arc material (Withnall and Lang, 1993), suggesting if an arc was present then it was probably proximal.
- Minor sodic I-type granitic magmatism in the Graveyard Creek Subprovince, most probably Ordovician in age (Withnall and Lang, 1993)

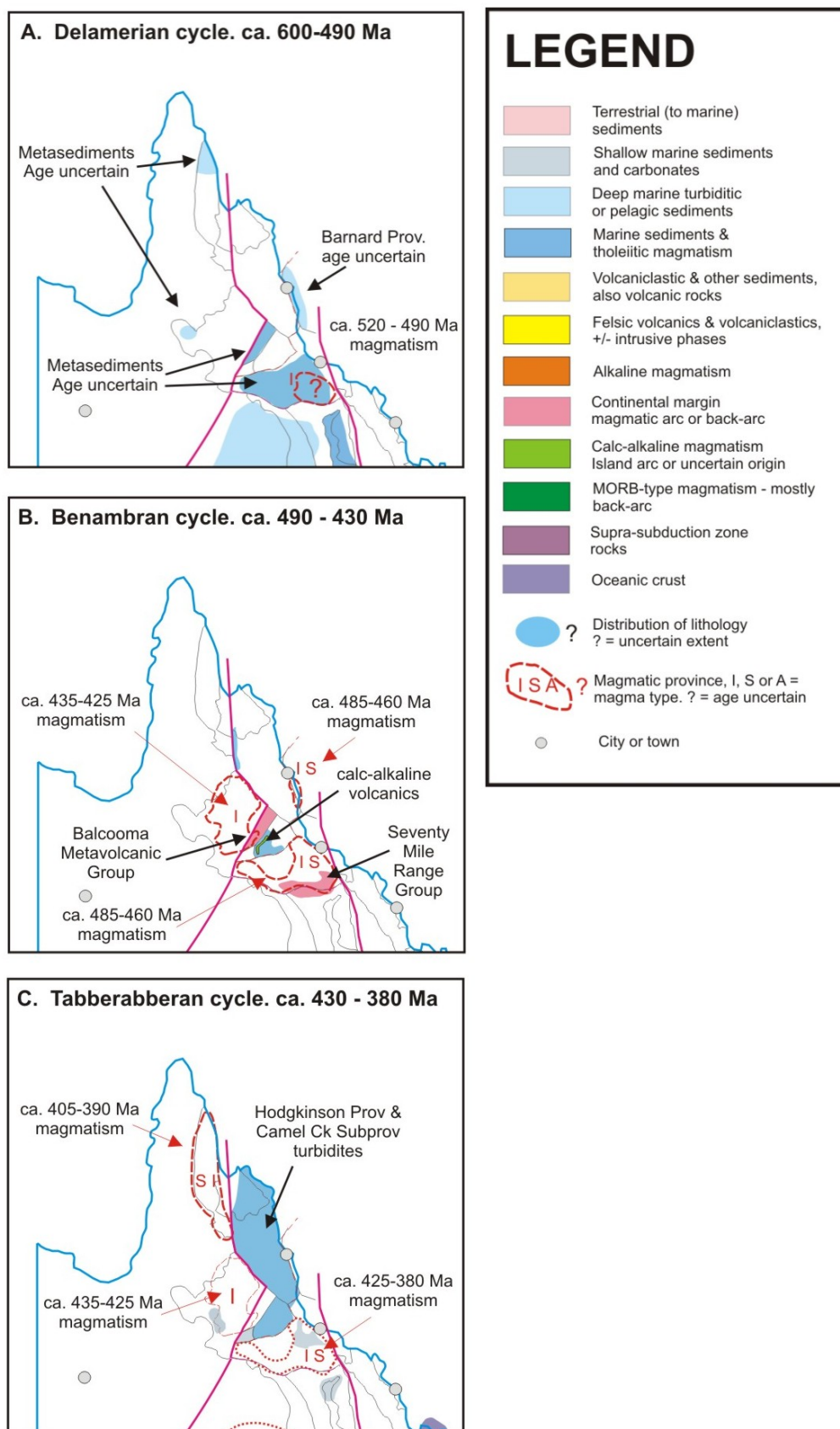


Figure 14. Generalised distribution of rocks in the North Queensland Orogen, by tectonic cycle. A = Delamerian cycle, B = Benambran cycle, C = Tabberabberan Cycle.

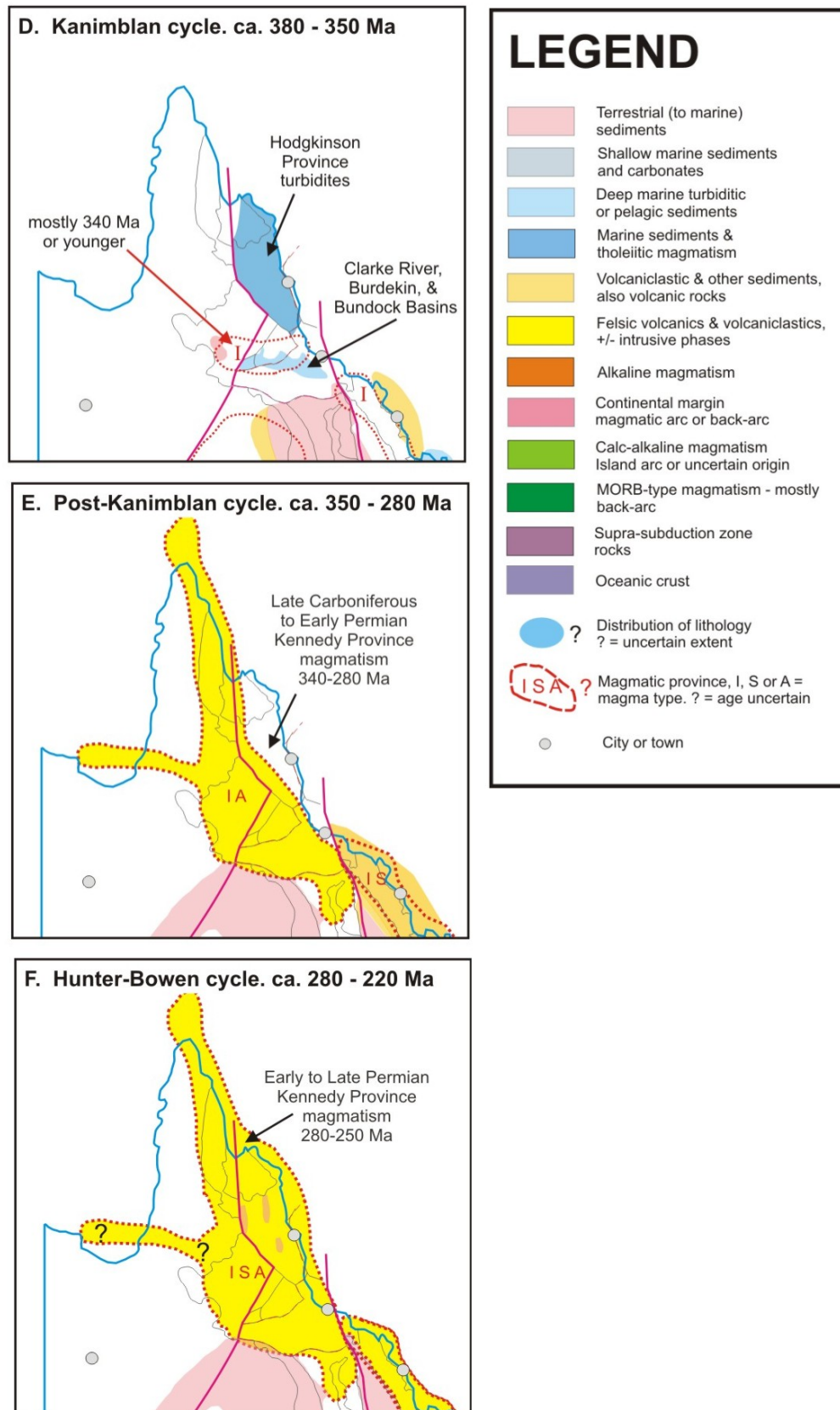


Figure 14 (continued). Generalised distribution of rocks in the North Queensland Orogen, by tectonic cycle. D = Kanimblan cycle, E = post-Kanimblan cycle, F = Hunter-Bowen cycle.

- As summarised by Arnold and Fawckner (1980) and Withnall and Lang (1993), the Graveyard Creek and Camel Creek subprovinces appear to have different structural histories (Fig. 13). The Judea Formation (Graveyard Creek Subprovince) records subhorizontal, mélange-type deformation and low grade metamorphism, that is no younger than Late Ordovician (Withnall and Lang, 1993; Withnall et al., 1997b) and may be syndeformational (e.g., Arnold and Fawckner, 1980). Withnall et al. (1997b) also record mélange-style deformation and low-grade metamorphism in the Camel Creek Subprovince. The age, however, is poorly constrained and Withnall et al. (1997b) suggested much of this deformation occurred in the Devonian. This deformation may equate with similar aged deformation in the Hodgkinson Province (Fig. 13).

1.2.4. Middle Silurian to Middle - early Late Devonian

Post-Benambran to Tabberabberan Orogeny: ca. 430 Ma to ca. 380 Ma

Geological and tectonic summary

This time period in north Queensland is characterised by extensive, probably related, sedimentation in the Hodgkinson and Broken River Provinces (along the eastern and southeastern margins of the Proterozoic Georgetown region), and also in the Charters Towers region (Fig. 14). Sedimentation includes marine siliciclastic sediments and carbonates, along with locally abundant, tholeiitic mafic volcanics (Arnold and Fawckner, 1980). Sediment provenance is dominantly cratonic (e.g., Bultitude et al., 1997) but does include volcanoclastic material, some of which is older (e.g., Garrad and Bultitude (1999) record dacitic clast ages of 465 Ma in a conglomerate from the Hodgkinson Province). The geodynamic environment for this sedimentation is controversial. Models include both backarc or forearc deposition, along with rifted continental margin (see summaries in Arnold and Fawckner, 1980; Garrad and Bultitude, 1999). Arnold (in Arnold and Fawckner, 1980), Henderson et al. (1980) and Henderson (1987), amongst others, suggested that the Hodgkinson and Broken River Province sedimentation was part of a forearc and accretionary wedge. This model, however, has difficulties explaining the tholeiitic volcanism, especially in the Hodgkinson Province. The latter is more consistent with rift or backarc models (e.g., Fawckner in Arnold and Fawckner, 1980; Bultitude et al., 1997), though could reflect accreted oceanic crust. Resolution of the geodynamic environment is critical to understanding the tectonics of the widespread Pama Province which comprises all the Silurian to Devonian magmatism in north Queensland (Bain and Draper, 1997). The Pama Province magmatism forms an extensive quasi-continuous belt around the Hodgkinson and Broken River Provinces, from Charters Towers in the south, north to Cape York (Fig. 14). In forearc models, this belt is interpreted as the magmatic arc (e.g., Henderson, 1987), although, as pointed out by numerous authors, the chemistry of this magmatism, especially in the Coen region, is not consistent with a magmatic arc (e.g., Blewett et al., 1997). Pama Province magmatism in the region is diachronous. Magmatic ages range from ca. 430-420 Ma (syn-Benambran and younger) in the Georgetown region, to ca. 425 to 405 Ma (and younger) in the Charters Towers region, to ca. 410 to 395 Ma in the Coen region. As pointed out by Champion and Bultitude (2003), these age differences are also matched by changes in geochemical signature. The early Pama magmatism (in the Georgetown region) does have geochemical signatures more consistent with arc magmatism. It is possible, therefore, that the Hodgkinson (and Broken River) Province was in a forearc environment early (ca. 430 Ma), and then evolved in a backarc environment (ca. 420-400 Ma and younger?).

The Tabberabberan Orogeny as defined in the Lachlan Orogen is constrained at ca. 385-375 Ma (e.g., VandenBerg et al., 2000; Gray and Foster, 2004; Glen, 2005); and appears to be just post-380 Ma (age of Mount Morgan Tonalite) in the New England Orogen (ca. 381 Ma; Golding et al., 1994; Yarrol Project Team, 1997, 2003). Deformations around this age in north Queensland are best represented in the Broken River Province, the Charters Towers region and possibly Hodgkinson Province, where they include time breaks and slight angular unconformities (Withnall and Lang, 1993; Fig. 13). Henderson (1987) suggested that this event resulted in the cessation of deep-marine sedimentation in the Camel Creek Subprovince, in

the Middle Devonian, and produced the angular unconformity observed in the Graveyard Creek Subprovince. Garrad and Bultitude (1999) record east to north-east thrusting and north-northwest-trending shear zones in the Hodgkinson Province, which they suggested were of Late Devonian age, possibly related to basin inversion. Although of a younger age, the latter may equate with the Tabberabberan Orogeny. Given the strong commonalities between the Broken River and Hodgkinson Provinces, it is probable that deep-water marine sedimentation ceased simultaneously in both. Arnold and Fawcner (1980) suggested syn-deformational *mélange* formation in the Hodgkinson province through the Tabberabberan Cycle.

There are also older deformation events, ca. 410-400 Ma, recorded in the Georgetown, Coen and Charters Towers regions (Fig. 13). The Coen and Charters Towers deformation coincides with Pama Province magmatism in those regions. Bultitude et al. (1997) record a change in sedimentation in the Hodgkinson Province in the Late Lochkovian (ca. 412 Ma). Similarly, Withnall et al. (1997b) record a hiatus in sedimentation at this time in the Graveyard Creek Subprovince. Both suggested these changes were related to hinterland uplift. Notably, sedimentation appears to recommence at this time in the Charters Towers region. These ca. 410-400 Ma events may correspond to the Tabberabberan Orogeny, though probably relate to the Bindian Orogeny. The latter is dated at ca. 420-410 Ma in the Lachlan Orogen (e.g., VandenBerg et al., 2000; Gray et al., 2003).

Geological history and Time-Space plot explanation

Georgetown (and Greenvale) region

- Isolated pockets of, largely Lower Devonian, marine, and locally terrestrial, sediments and limestone (Withnall et al., 1997b).
- East-west shortening and greenschist metamorphism (Withnall et al., 1997a), tentatively dated at ca. 400 Ma, but may form part of the 430 Ma Benambran event (Fig. 13). This age does, however, coincide with lode Au mineralisation (410-400 Ma; Withnall et al., 1997a), and with extensive magmatism to the north in the Coen region.

Barnard region

- Regional deformation and associated low to potentially high-grade metamorphism (Bultitude et al., 1997). This poorly constrained deformation event is bracketed between ca. 460 Ma and Permian (Garrad and Bultitude, 1999).

Coen region

- Voluminous, S-type-dominated, mostly felsic, magmatism of the Pama Province. Dominated by the S-type Kintore Supersuite (Fig. 14). Ages constrained between ca. 410 Ma and 395 Ma (Blewett et al., 1997).
- East-west shortening deformation and low-P high-T metamorphism (to upper amphibolite-facies) with some contemporaneous Pama Province magmatism – ca. 407 Ma (Blewett et al., 1997).
- East-west shortening and thrusting, which Blewett et al. (1997) equated with inversion of the Hodgkinson Province (Fig. 13).

Charters Towers region

- Largely mixed siliciclastic marine sediments and carbonate successions of the Wilkie Gray and Fanning River Groups (Hutton et al., 1997; Figs 13, 14). Although this sedimentation essentially continues through to the Early Carboniferous, there is an unconformable contact and a significant change in sedimentation above the Fanning River Group (Hutton et al., 1997).
- Extensive magmatic rocks of the Pama Province. In the Charters Towers region, these include the widespread I-type Millchester Supersuite in the Ravenswood Batholith as well as I-types in the Reedy Springs Batholith. Ages are best constrained in the Ravenswood Batholith, where they range from ca. 425 Ma to ca. 405 Ma (Hutton et al., 1997). Similar ages appear to occur in the

Reedy Springs and Lolworth batholiths, although the latter also include widespread younger (ca. 380 Ma) granites which include S-types (Hutton et al., 1997).

- Northeast to east-northeast faulting, with associated lode gold vein formation, ca. 400 Ma (e.g., Hutton et al., 1997).

Hodgkinson region

- Early Silurian to Early Devonian limestone, tholeiitic basalt and siliciclastic marine turbiditic sediments of the Chillagoe Formation (Arnold and Fawckner, 1980; Garrad and Bultitude, 1999; [Fig. 14](#)).
- Extensive marine, largely siliciclastic, turbiditic sediments – largely in the Hodgkinson Formation (Arnold and Fawckner, 1980; Bultitude et al., 1997; Garrad and Bultitude, 1999; [Fig. 14](#)). The latter also includes minor tholeiitic metabasalts (Arnold and Fawckner, 1980). Upper and lower age limits not well constrained but at least Early to Late Devonian in age (Garrad and Bultitude, 1999).
- No unequivocal deformation of Tabberabberan age appears to have affected the Hodgkinson Province as sedimentation continues throughout the Late Silurian and Devonian, although Arnold and Fawckner (1980) suggested syn-deformational *mélange* formation. Also, Garrad and Bultitude (1999) record east to north-east thrusting and north-northwest-trending shear zones in the Hodgkinson Province, which they suggested were of Late Devonian age, possibly related to basin inversion, and possibly related to the Tabberabberan Orogeny.

Camel Creek and Graveyard Creek regions

- In the Camel Creek Subprovince, continuation of marine, quartzose, turbiditic sedimentation, locally with tholeiitic metabasalt and limestone, possibly to as young as Early Devonian (Arnold and Fawckner, 1980; Withnall and Lang, 1993; [Fig. 14](#)).
- Largely mixed siliciclastic marine (to locally fluvial) sediments and carbonate successions (Withnall and Lang, 1993; Withnall et al., 1997b) in the Graveyard Creek Subprovince ([Fig. 14](#)). This sedimentation appears to have continued through the Silurian to the Early Devonian, although Withnall and Lang (1993) record time breaks between the Graveyard Creek Group and Shield Creek Formation and Broken River Group (Lochkovian and early Emsian; [Fig. 13](#)).
- Southeast-northwest deformation, and accompanying greenschist to amphibolites-facies metamorphism, is recorded in the Graveyard Creek Subprovince (e.g., Arnold and Fawckner, 1980). This deformation produced mild angular unconformities, and is constrained to be older than Early Frasnian (Withnall et al., 1997b; [Fig. 13](#)). The Camel Creek Subprovince records two events – east-directed thrusting and folding, and southeast-northwest deformation with accompanying greenschist-facies metamorphism (Arnold and Fawckner, 1980; Withnall and Lang, 1993). As suggested by Withnall et al. (1997b), ages for these events are constrained between Early Devonian and Early Carboniferous, but both deformations may equate with that recorded in the Graveyard Creek Subprovince. Arnold and Fawckner (1980) suggested this deformation was Late Devonian in age. Henderson (1987) suggested that the southeast-northwest deformation affected both subprovinces, shutting off deep marine sedimentation in the Camel Creek Subprovince, and producing the angular unconformity in the Graveyard Creek Subprovince.

1.2.5. Late Devonian to Early Carboniferous

Post-Tabberabberan to Kanimblan Orogeny: ca. 380 Ma – ca. 350 Ma

Geological and tectonic summary

This time period in north Queensland produced largely non-volcanic (cratonic provenance), terrestrial and lesser marine sedimentation across all regions, best preserved in the lower successions of the Bundock, Clarke River and Burdekin basins of the Broken River and Charters Towers regions (Fig. 14). Minor andesitic volcanism is recorded in the Georgetown region (Withnall et al., 1997a), and minor volcanoclastic input is recorded in several of the regions. Subsequent deformation is minor in nearly all areas, with the exception of the Hodgkinson Province where significant east-west shortening and further basin inversion occurred (Garrad and Bultitude, 1997; Fig. 13). This deformation and time period immediately pre-dates the commencement of the voluminous and very widespread extrusive and intrusive magmatism of the Kennedy Province. Tectonic environment for the Kanimblan cycle in north Queensland is poorly constrained. The possible continuation of largely Tabberabberan Cycle turbiditic sedimentation in the Hodgkinson Province suggests a similar geodynamic regime to the previous cycle, i.e., deposition in either a backarc or forearc environment (e.g., Arnold and Fawckner, 1980; Garrad and Bultitude, 1999; see Section 1.2.4).

Geological history and Time-Space plot explanation

Georgetown region (and Greenvale Subprovince)

- Sporadic belts and blocks of terrestrial and marine sediments (Withnall et al., 1997a; Withnall and Lang, 1993). Rocks are mostly non-volcanic but do include, locally abundant, andesite lavas (Withnall et al., 1997a).
- Poorly constrained, very open, north-south deformation, thought to be mid Carboniferous, pre ca. 335 Ma (Withnall et al., 1997a).

Barnard Province

- Regional deformation and associated low to potentially high-grade metamorphism (Bultitude et al., 1997). This poorly constrained deformation event is bracketed between ca. 460 Ma and Permian (Garrad and Bultitude, 1999).

Coen Region

- Widespread terrestrial sediments ($>1000 \text{ km}^2$), including coal, in the Pascoe River Beds (Blewett et al., 1997; McConachie et al., 1997). Volcanic detritus has been recorded in the succession, and it appears to include a lower volcanic unit of uncertain age (McConachie et al., 1997).
- Poorly constrained minor deformation, thought to be Carboniferous and/or Permian in age (Blewett et al., 1997). McConachie et al. (1997) indicate that the Pascoe River Beds are deformed, suggesting that deformation was post-Early Carboniferous, and pre-dated granite intrusion (ca. 285 Ma).

Charters Towers Region

- Terrestrial and marine sedimentation, including volcanoclastics in the Dotswood and Keelbottom Groups, with a number of transgressive-regressive cycles in the upper part (Early Carboniferous), e.g., Hutton et al. (1997).
- No apparent deformation recorded.

Hodgkinson Province

- Continuation of the extensive marine, largely siliciclastic, turbiditic sediments of the Hodgkinson Formation (Arnold and Fawckner, 1980; Bultitude et al., 1997; Garrad and Bultitude, 1999; Fig.

- 14). Upper age limits not well constrained but at least Late Devonian, and possibly very earliest Carboniferous in age (Garra and Bultitude, 1999).
- Locally distributed marine and terrestrial to marginal marine sediments, thought to be mostly Early Carboniferous in age (Garra and Bultitude, 1999).
 - Significant early to mid Carboniferous deformation (predates Late Carboniferous magmatism), which according to Davis (1994) and Garra and Bultitude (1999) produced significant east-west shortening. Bultitude et al. (1997) included north-south shortening during this period. Garra and Bultitude (1999) suggested the latter was Permian in age. The north-south orientation, however, suggests the deformation is probably Carboniferous and equates to the Alice Springs Orogeny.

Broken River Province

- In both subprovinces, there are well developed successions of terrestrial and lesser marine sedimentation (Fig. 14). These are mostly of cratonic provenance, though do have some volcanoclastic input (Withnall and Lang, 1993, Withnall et al., 1993). This style of sedimentation changes in the Visian with the commencement of widespread volcanism.
- No deformation recorded but there are apparent time breaks which coincide with a significant change in sedimentation style, namely the onset of regional Kennedy Province magmatism (intrusive and extrusive) in the Visian.

1.2.6. Middle Carboniferous to Latest Permian

Post-Kanimblan to Hunter-Bowen Orogeny: ca. 350 Ma to 250 Ma

Geological and tectonic summary

This period in north Queensland is characterised by the commencement of the widespread and voluminous extrusive and intrusive Kennedy province magmatism, plus associated, mostly minor sedimentation (Fig. 14). As documented by numerous authors (e.g., Richards et al., 1966) this magmatism is crudely diachronous, commencing earlier in the Georgetown, Broken River and Charters Towers regions (Figs 13, 14), in the Visian (ca. 340-335 Ma), and younging to mid Permian in the Hodgkinson Province. There are also accompanying changes in geochemistry. Magmatism in the mid to Late Carboniferous is almost exclusively I-type in nature, with some mantle-derived magmatism. In the Early Permian, magmatism switched to A- and I-type in the Georgetown and western Hodgkinson regions, and to S- and I-type in the central and eastern Hodgkinson (Champion and Bultitude, 2003).

There are at least two deformations during this time period. There is a widespread but minor north-south contraction, thought to be mid to Late Carboniferous in age, and commonly equated to the Alice Springs Orogeny, found in the Broken River, Hodgkinson and Georgetown regions (Withnall et al., 1997a, b; Bultitude et al., 1997; Fig. 13). In addition, there is a more significant deformation in the Early Permian. This penetrative east-west shortening deformation is best documented in the Hodgkinson Province (e.g., Davis, 1994; Davis et al., 1996; Garra and Bultitude, 1999; Fig. 13). The Late Permian is characterised by east-west deformation, especially in the Hodgkinson and Barnard provinces (Garra and Bultitude, 1999). This deformation is equated with the Hunter-Bowen Orogeny.

The tectonic regime for this region is not well understood. Magmatism of the Kennedy Province, although dominated by crustal input, is generally thought to be broadly arc-related, possibly back-arc (e.g., Champion and Bultitude, 2003). This is consistent with the New England Orogen, which at this time, was characterised by a convergent margin environment, alternating between extension and contraction (e.g., Van Noord, 1999).

Geological history and Time-Space plot explanation

Georgetown (and Greenvale) region

- Mid-Carboniferous to Early Permian voluminous extrusive and intrusive magmatism of the Kennedy Province. Magmatism ranges in age from ca. 340 Ma to ca. 275 Ma (e.g., Withnall et al., 1997a; Fig. 14). It is mostly felsic, but includes minor mafic and intermediate products. Magmatism is dominantly I-type but includes widespread but generally small volume A-types (e.g., Champion and Bultitude, 2003). The I-type magmatism shows a broad decrease in age from ca. 340 Ma in the southwest to ca. 280 Ma in the northeast, e.g., Richards et al. (1966), Champion and Bultitude (2003). The A-types, however, are, where dated, typically younger, ca. 290-275 Ma (Withnall et al., 1997a; Champion and Bultitude, 2003), and appear to occur across the region. Magmatism of this age is extensively mineralised, mostly associated with the I-type granites.
- Minor, dominantly terrestrial, sedimentation is associated with volcanic complexes.
- Poorly defined, east-west deformation, thought to be Late Permian in age and equated with the Hunter-Bowen Orogeny (Withnall et al., 1997a).

Barnard Province

- Permian S-type, and less common I-type, Kennedy Province magmatism (Bultitude et al., 1997; Garrad and Bultitude, 1999).
- Regional deformation and associated low (to high?) grade metamorphism (Bultitude et al., 1997) related to the Hunter-Bowen Orogeny. This event strongly deforms the Early Permian granites and so must postdate ca. 280 Ma (Garrad and Bultitude, 1999).

Coen Region

- Late Carboniferous to Early Permian Kennedy Province magmatism (extrusive and intrusive). Contains (intermediate to) felsic I- and A-types, from ca. 310 Ma (in the Torres Straits; Von Gnielinski et al., 1997) to Middle Permian (ca. 275 Ma, Blewett et al., 1997; Fig. 14).
- Permian terrestrial sediments, including coal, in the concealed Olive River Basin (McConachie et al., 1997).
- Poorly constrained Carboniferous to Permian(?) deformation and low grade metamorphism (Blewett et al., 1997). Both east-west and younger north-south deformations occur. The latter is most consistent with events equated to the Alice Springs Orogeny, and as such is probably early-Mid Carboniferous in age. Younger events may be present; certainly McConachie et al. (1997) record unconformities above and below the Olive River Basin.

Charters Towers region

- Terrestrial sedimentation and basaltic to felsic volcanics of the Early to mid Carboniferous Glenrock Group (Hutton et al., 1997). Of similar age (ca. 340-330 Ma) are co-magmatic rhyolitic volcanism and felsic granites of the Oweenee Supersuite (Hutton et al., 1997), as well as other unrelated granites.
- Late Carboniferous to Early Permian felsic volcanism of the Kennedy Province and terrestrial sediments (Fig. 14). These rocks appear to unconformably overlie the Early to mid Carboniferous Glenrock Group and other similar aged volcanics.
- Late Carboniferous to mid Permian mainly felsic I-type granites and associated (minor) mafic intrusives, of the Kennedy Province. Hutton et al. (1997) indicated magmatism broadly youngs in age from ca. 310 Ma in the west to as young as ca. 265 Ma in the east. Younger magmatism has transitional A-type characteristics.
- Hutton et al. (1997) suggested a relatively stable backarc setting for the Kennedy Province magmatism with a dominant extensional environment throughout this period. The recognition of a potential unconformity between the Early to mid Carboniferous volcanics and overlying rocks may indicate some Alice Springs Orogeny effects.

Hodgkinson Province

- Extensive I- and A-type Kennedy Province extrusive and intrusive magmatism, and minor terrestrial sediments, concentrated in the western and southern part of the Hodgkinson Province (Champion and Bultitude, 2003; Fig. 14). The I-types have the greatest age range - ca. 320 Ma to ca. 275 Ma (Garrad and Bultitude, 1999). The A-types are Permian in age (ca. 290 to 275 Ma; Garrad and Bultitude, 1999). The I-type intrusives, in particular are extensively mineralised.
- Permian S-type, and less common I-type, Kennedy Province magmatism in the eastern half of the Hodgkinson Province (Bultitude and Champion, 1992; Champion and Bultitude, 1994; 2003; Fig. 14). Magmatism appears to young to the northeast and east, from ca. 280 Ma to ca. 260 Ma and potentially younger (Richards et al., 1966; Garrad and Bultitude, 1999).
- Locally distributed, largely terrestrial sediments and coal measures of Late Permian age (Garrad and Bultitude, 1999).
- Regional (syn-granite emplacement) deformation (ca. 275-280 Ma; Fig. 13), best documented by Davis (1994). According to Davis (1994) this deformation, which produced north-south fabrics, was strongly partitioned and so is variably developed across the Province. This age corresponds closely to that of A-type magmatism in the western Hodgkinson Province and Georgetown region, which is commonly interpreted as being emplaced in an extensional environment.
- Significant Late Permian to Early Triassic transpressive(?) deformation and greenschist metamorphism thought to be related to the Hunter-Bowen Orogeny (Garrad and Bultitude, 1999; Fig. 13). Bultitude et al (1997) document K-Ar age resetting in granites ca. 250 Ma in age.
- Minor Triassic terrestrial sediments, which, at least locally, overlie older Permian rocks with an angular unconformity (Garrad and Bultitude, 1999).

Broken River Province (Withnall and Lang, 1993; Withnall et al., 1997b)

- Viséan and younger terrestrial sedimentation and felsic volcanics in the upper parts of the Bundock and Clarke River Groups. These rocks equate with the Glenrock Group in the Charters Towers region, though lack the mafic-intermediate rocks found in the latter.
- The margins of the province shares similar Kennedy magmatism to that documented for the Hodgkinson and Charters Towers regions.

1.3. New England Orogen

by N Kositsin

Introduction

The New England Orogen (NEO, [Figs 15, 16](#)) is a complex collage of different terranes on the eastern margin of the Australian continent (Cawood and Leitch, 1985; Leitch and Scheibner, 1987; Flood and Aitchison, 1988). The NEO has had a complicated evolutionary history that stretches from the Neoproterozoic to the Late Mesozoic, although most of the terranes are Silurian to Carboniferous in age (Aitchison et al., 1992a; Aitchison and Ireland, 1995). The major component of the orogen evolved during the Devonian and Carboniferous in a convergent plate margin tectonic setting related to a west-dipping subduction system (Murray et al., 1987; Korsch et al., 1990; Fergusson et al., 1993).

Parts of the volcanic arc, forearc basin and accretionary wedge are still preserved in the orogen ([Fig. 16](#)). In the northern NEO, Devonian and Carboniferous rocks can be subdivided into arc, forearc, and accretionary wedge assemblages, consisting of the Connors and Auburn Arches in the west (magmatic arc), the Yarrol Belt in the centre (forearc basin), the accretionary wedge to the east (Coastal, Yarraman, North D'Aguilar, South D'Aguilar and Beenleigh terranes) (Murray, 1986; [Figs 15, 16](#)), and a backarc basin (Drummond Basin). In the southern NEO, the arc is inferred to have been located to the west of the Tamworth Belt (either concealed beneath the Gunnedah Basin or Tamworth Belt, or removed by erosion and/or strike-slip faulting; Korsch et al., 1997). The forearc basin is represented by the Tamworth Belt and Hastings Block, and the accretionary wedge by the Tablelands Complex (Woolomin, Sandon and Coffs Harbour Associations of Korsch, 1977, or in this compilation, the Woolomin, Wisemans Arm, Central and Coffs Harbour terranes; [Fig. 15](#)). The present boundary between forearc basin strata to the west and subduction complex assemblages to the east is a major fault zone marked by serpentinite lenses (the Yarrol Fault in the northern NEO and the Peel-Manning Fault System in the southern NEO; Murray, 1988; [Fig. 15](#)). The Gympie Province, in the northern NEO, is thought to be an exotic terrane (e.g., Korsch and Harrington 1987; Murray, 1988).

Numerous tectonic models have been proposed for the development of the NEO (e.g., Leitch, 1975; Cawood 1982, 1983; Murray et al., 1987). Most infer a long-lived, east-facing convergent plate margin setting, with progressive accretion of younger rocks at the eastern margin of Gondwanaland. The orogen was most likely island arc-related from the Cambrian to the Early Devonian, evolving to a continental margin magmatic arc from the Late Devonian onwards. The subsequent history of the orogen involved strike-slip faulting and major oroclinal bending. Large amounts of volcanism and plutonism took place in the Late Permian and Early Triassic (e.g., Korsch et al., 1990).

A brief synthesis, highlighting key points in the evolution of the NEO from the Neoproterozoic through to the Triassic, is presented in the following sections and accompanying time-space and other plots. The time-space plots ([Figs 17, 18, 19](#)) summarise our current state of knowledge of the stratigraphic history ([Fig. 17](#)) and tectonic evolution ([Fig. 18](#)) of the NEO through time.

1.3.1. Late Neoproterozoic to earliest Ordovician

Rodinian break-up to Delamerian Orogeny: ca. 520 to ca. 490 Ma

The New England Orogen in this period is dominated by Cambrian tectonic blocks, largely of oceanic fragments, including island arc-related remnants ([Fig. 19](#)). These occur in the southern NEO, as accreted blocks along the Peel-Manning Fault System (PMFS, e.g., Offler and Shaw, 2006; [Fig. 15](#)). These blocks provide records of subduction environments offshore of continental Australia in the (latest Neoproterozoic) Cambrian. They were accreted to the NEO in the Late Palaeozoic.

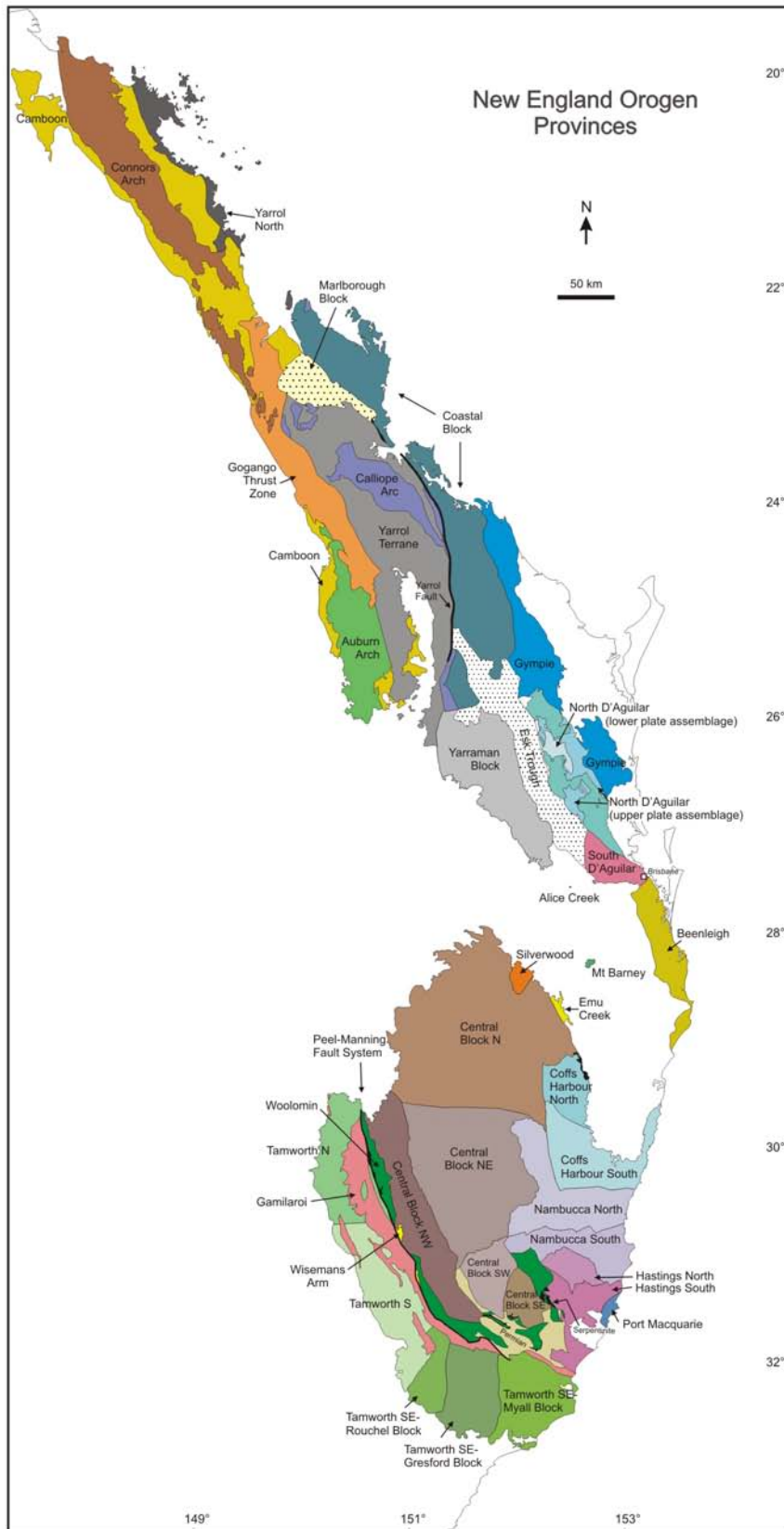


Figure 15. Distribution of geological provinces and blocks in the New England Orogen used to construct the time-space plots.

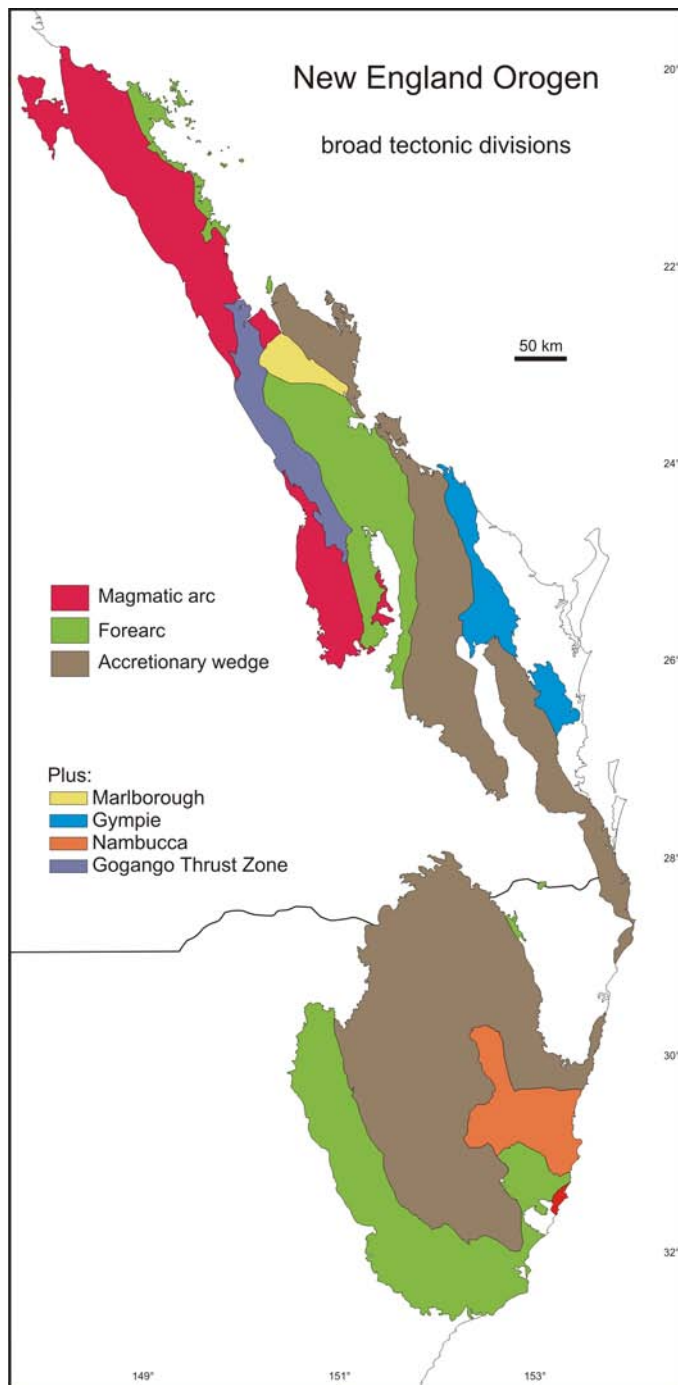


Figure 16. Broad subdivision of the New England Orogen into regions defined by tectonic setting.

The only definitive record of Neoproterozoic-Cambrian material in the northern NEO is represented by the Princhester and related ophiolites, which are ca. 565 Ma in age (Bruce et al., 2000; Fig. 19). These were most probably accreted to the NEO in the Late Palaeozoic (Murray and Blake, 2005).

Southern NEO

- In the southern New England Orogen, the Peel-Manning Fault System (PMFS; Fig. 15) is a major structural element characterised by tectonic blocks in serpentinite mélangé (Offler and Shaw, 2006).
- The PMFS contains fragments of dismembered ophiolite of suprasubduction origin (Aitchison et al., 1994), the oldest of which is Cambrian in age. Eclogite blocks (Attunga Eclogite), exhumed along the PMFS were originally dated at 571 Ma by Watanabe et al. (1998); this date has now been revised to ca. 536 Ma (Fanning et al., 2002). Island arc boninitic volcanism at 536 Ma associated with subduction and high P-low T metamorphism of MORB-like basalts to eclogite, therefore, commenced in the Early Cambrian. This record of Cambrian subduction is earlier than boninitic magmatism recorded in Tasmania (514±5 Ma; Black et al., 1997) and Victoria (519-514 Ma; VandenBerg et al., 2000).
- Similar ages to the Attunga Eclogite have been obtained from plagiogranite (U-Pb zircon 530±6 Ma; Aitchison et al., 1992a) and metadiorite (Sm-Nd isochron 536±22 Ma; Sano et al., 2004) in schistose serpentinite along the same fault system. The plagiogranite age suggests 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 and Ireland, 1995). Sano et al. (2004) suggest that the boninitic metadiorite formed in an immature island arc setting as a result of mixing of a depleted mantle wedge with sediment melt and fluid and melt derived from MORB.
- The Rocky Beach Metamorphic Mélange (ca. 545 Ma) and the Port Macquarie Serpentinite (~545-509 Ma) are inferred to be the oldest rocks in the Port Macquarie Block (Figs 17, 19), largely dated by analogy with rocks from elsewhere in the NEO (e.g., Och et al., 2007). They contain a fragmentary late Neoproterozoic to early Palaeozoic history that includes subduction and associated metamorphism under high-P low T conditions, possibly around 536 Ma (Watanabe et al., 1998; Och et al., 2007). The Port Macquarie Serpentinite is a product of alteration of cumulate ultramafic rocks of a c. 530 Ma forearc ophiolite (Och et al., 2007).
- Palaeontological studies have revealed island arc-related activity during the Middle or early Late Cambrian in the Gamilaroi Terrane (Cawood, 1976; Leitch and Cawood, 1987; Stewart, 1995). Volcaniclastic rocks from the Murrawong Creek and Pipeclay Creek Formations contain Cambrian trilobite, brachiopod and conodont faunas (Cawood, 1976; Cawood and Leitch 1985; Stewart, 1995), and possibly form an ancient basement to the predominantly mid-Palaeozoic rocks of the Gamilaroi Terrane (Stratford and Aitchison, 1997). Cambrian conodonts in the Pipeclay Creek Formation (Stewart, 1995) indicate the presence of Cambrian sedimentary rocks below the main Devonian-Carboniferous pile in the Tamworth Belt, supporting the idea of Korsch et al. (1997) that perhaps at least some of the ophiolites were the floor to the Tamworth Belt. Cawood and Leitch (1985) have suggested that volcanic clasts in these sediments were derived from a low-K intra-oceanic island arc.

Northern NEO

- In the northern NEO, the rift phase of the Delamerian cycle is represented by an ophiolite (Northumberland and Princhester serpentinites) in the Marlborough Terrane of central Queensland that has a 562±22 Ma Sm-Nd isochron age (Bruce et al., 2000; Fig. 19). The ophiolite has depleted MORB-like trace element characteristics that suggest formation as oceanic crust at a Neoproterozoic ocean spreading ridge (Bruce et al., 2000). These data are consistent with the existence of a proto-Pacific Ocean east of the Delamerian Orogen after supercontinent breakup and

is consistent with the inferred presence of old lithosphere under the southern NEO (Glen, 2005). The Marlborough Terrane was emplaced into its present position during the Hunter-Bowen Orogeny (Holcombe et al., 1997b; Korsch et al., 1997; Murray and Blake, 2005).

- There appears to be no record in the northern NEO of supra-subduction zone ophiolites and volcanic arc deposits of Cambrian age, similar to those exposed along the PMFS. Age and geochemical data suggest that the ~562 Ma magmatism in the Princhester Serpentine must have occurred outboard of any Early Palaeozoic subduction zone along the margin of the Australian continent (Murray, 2007).

1.3.2. Earliest Ordovician-to earliest Silurian

Post-Delamerian Orogeny to Benambran Orogeny: ca. 490 to 440 Ma

The Benambran cycle geology in the New England Orogen is dominated by Ordovician and Silurian sedimentary rocks of island arc and oceanic affinity and Late Ordovician arc magmatism, the latter recorded along the PMFS (e.g., Offler and Shaw, 2004; Fig. 19). Early and Middle Ordovician blueschist metamorphism resulted from the exhumation of Cambrian to Early Ordovician rocks at Port Macquarie and along the PMFS.

- Following island arc boninitic volcanism at 536 Ma, exhumation of MORB-like basalts in the Early Ordovician gave rise to: high-pressure blueschist metamorphism (K/Ar ages of 482-467 Ma for blueschist rocks within serpentinite-matrix mélange at Glenrock and Pigna Barney; Fukui et al., 1995); arc-related plutonism, e.g., the Attunga gabbro (U-Pb zircon; 479 ± 11 Ma; Fanning et al., 2002); and accretionary wedge-related blueschist metamorphism (Fukui et al., 1995; Offler, 1999), along the PMFS (Offler, 1999). Offler and Shaw (2006) also provide evidence for Late Ordovician arc magmatism along the PMFS (U-Pb zircon age of 444.7 ± 2.4 Ma for hornblende gabbro of calcalkaline affinity at Glenrock station; see also Watanabe et al., 1998). Similarly, a U-Pb zircon age of 436 ± 9 Ma for tonalite of the Pola Fogal Suite in the Pigna Barney Ophiolite Complex (Kimbrough et al., 1993) may also be related to an Ordovician arc, though may also reflect Pb-loss from the Cambrian arc (Kimbrough et al., 1993).
- The oldest sedimentary rocks in the southern NEO are siliciclastic sedimentary rocks and fossiliferous limestones found within the imbricate zone of the PMFS, and include the Trelawney beds and Haedon Formation in the Gamilaroi Terrane (island arc related environment), and Uralba beds (probably part of Gamilaroi), preserved along the PMFS (Fig. 19). Corals and conodonts from these limestones indicate they are of Late Ordovician age (Hall, 1975). A similar-aged limestone succession lies below the Silverwood Group (Wass and Dennis, 1977; Silverwood Terrane; Figures 17 and 18). Late Ordovician coral limestone, probably representing partly accreted seamounts, was also deposited in an ocean basin environment, now incorporated in the Woolomin Terrane (435-428 Ma; Hall, 1978).
- Middle to Late Ordovician rocks (chert, mudstone, siltstone, tuffaceous sandstone, tuff, conglomerate, olistostromal rocks and basalt) of the Watonga Formation at Port Macquarie have been interpreted to have been deposited on oceanic crust prior to accretion (Och et al., 2007).
- K-Ar ages of c. 469 Ma for phengite related to retrograde blueschist metamorphism of eclogite blocks in the Port Macquarie Block has been interpreted to suggest that substantial exhumation of these rocks had occurred by the Middle Ordovician (Fukui, 1991; Watanabe et al., 1993; Fukui et al., 1995; Offler, 1999). In addition, glaucophane from blueschist in mélange of the Port Macquarie Block has a K-Ar isotopic age of 444 Ma (Lanphere, cited in Scheibner, 1985).

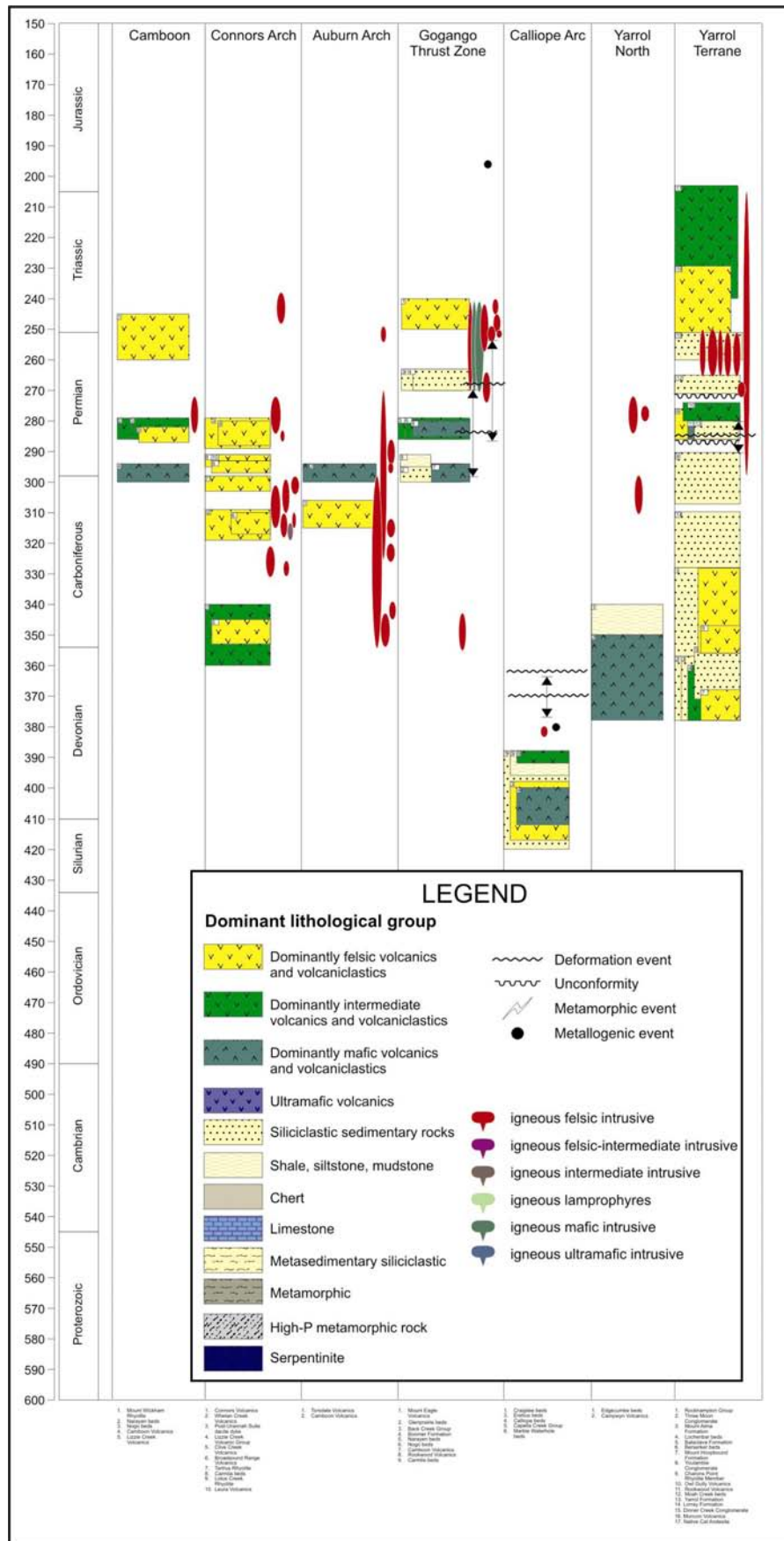


Figure 17. Late Neoproterozoic to Jurassic time-space plot for part of the northern New England Orogen. Refer to text for data sources and discussion. Terranes are as outlined in Figs 15 and 16.

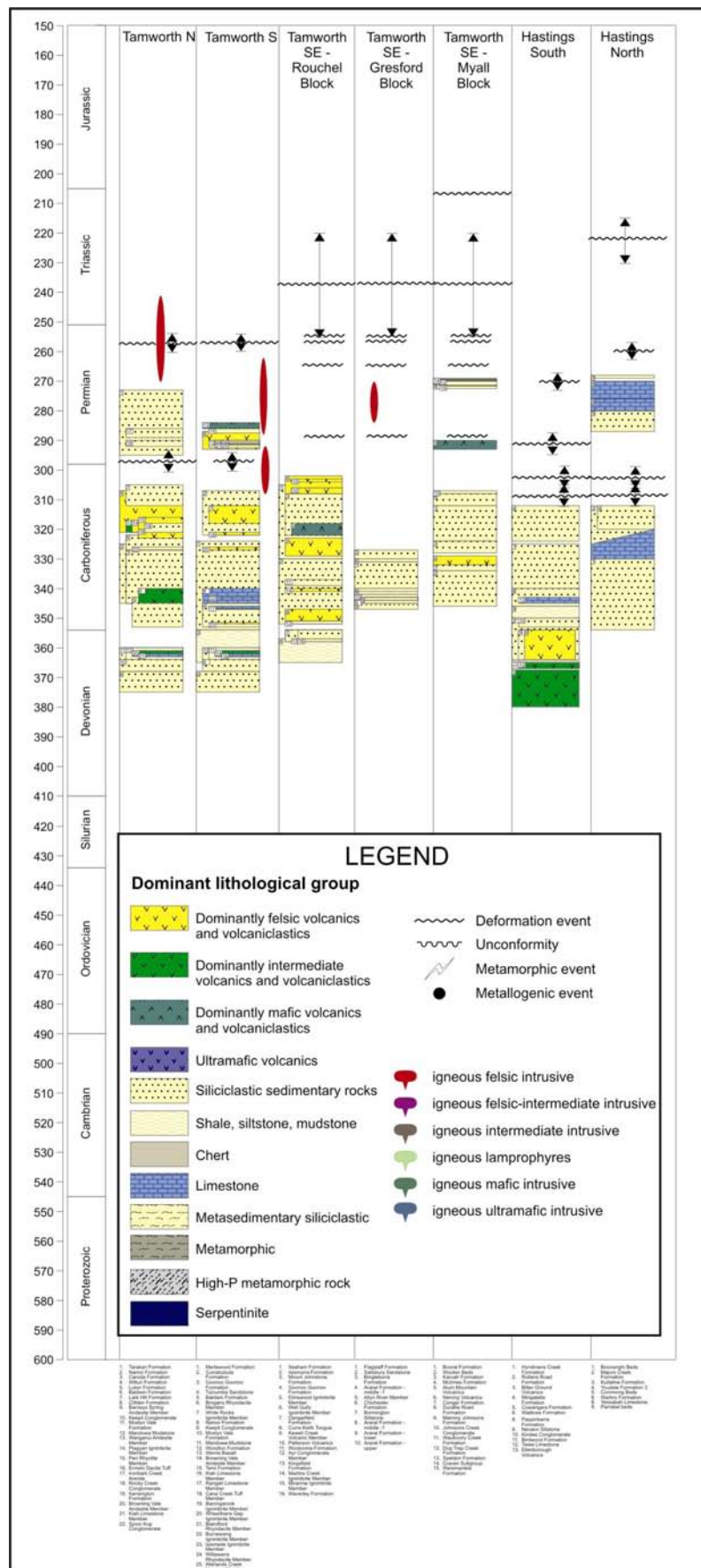


Figure 17 continued. Late Neoproterozoic to Jurassic time-space plot for part of the southern New England Orogen. Refer to text for data sources and discussion. Terranes are as outlined in Figs 15 and 16.

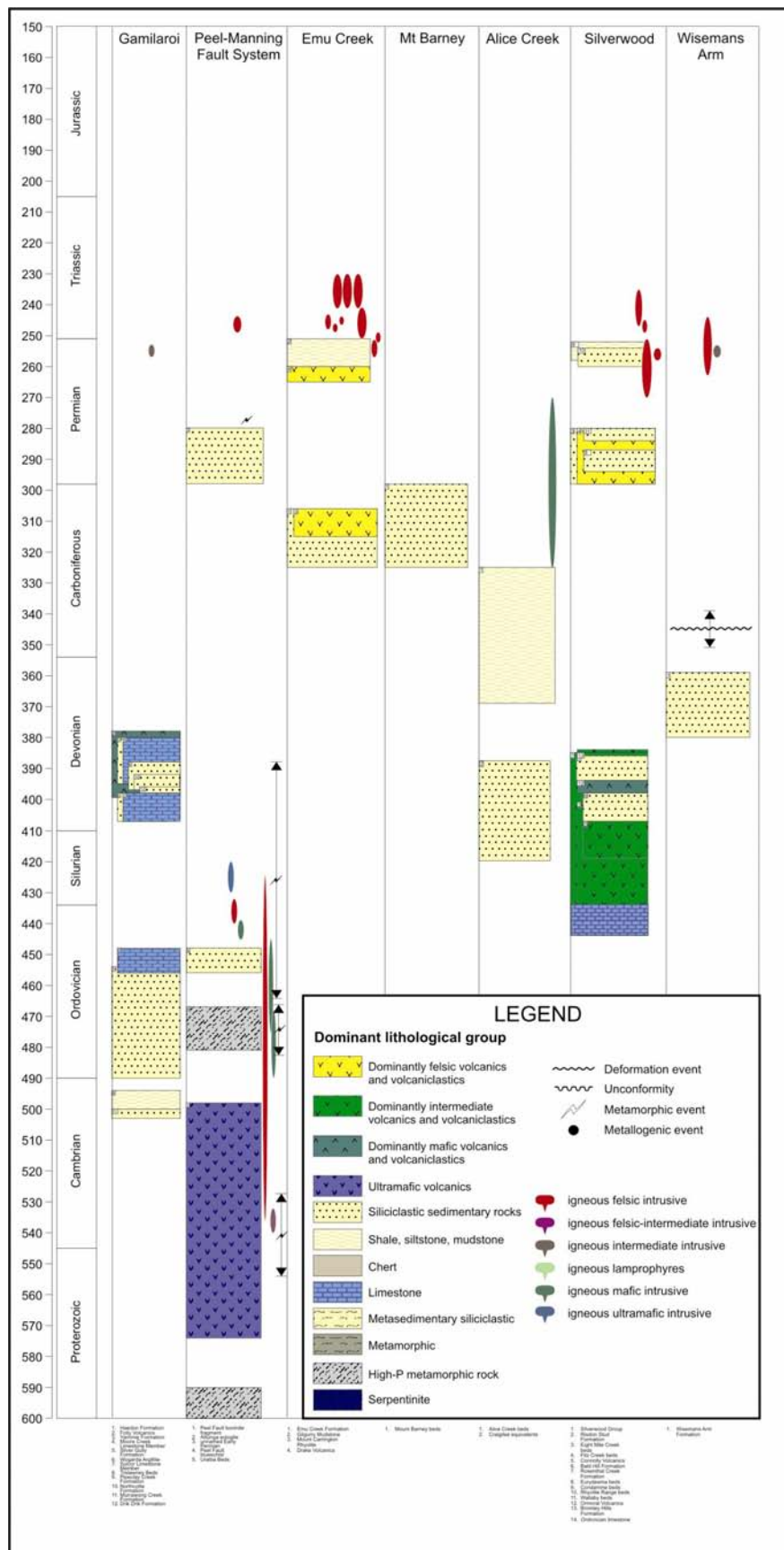


Figure 17 continued. Late Neoproterozoic to Jurassic time-space plot for part of the southern New England Orogen. Refer to text for data sources and discussion. Terranes are as outlined in Figs 15 and 16.

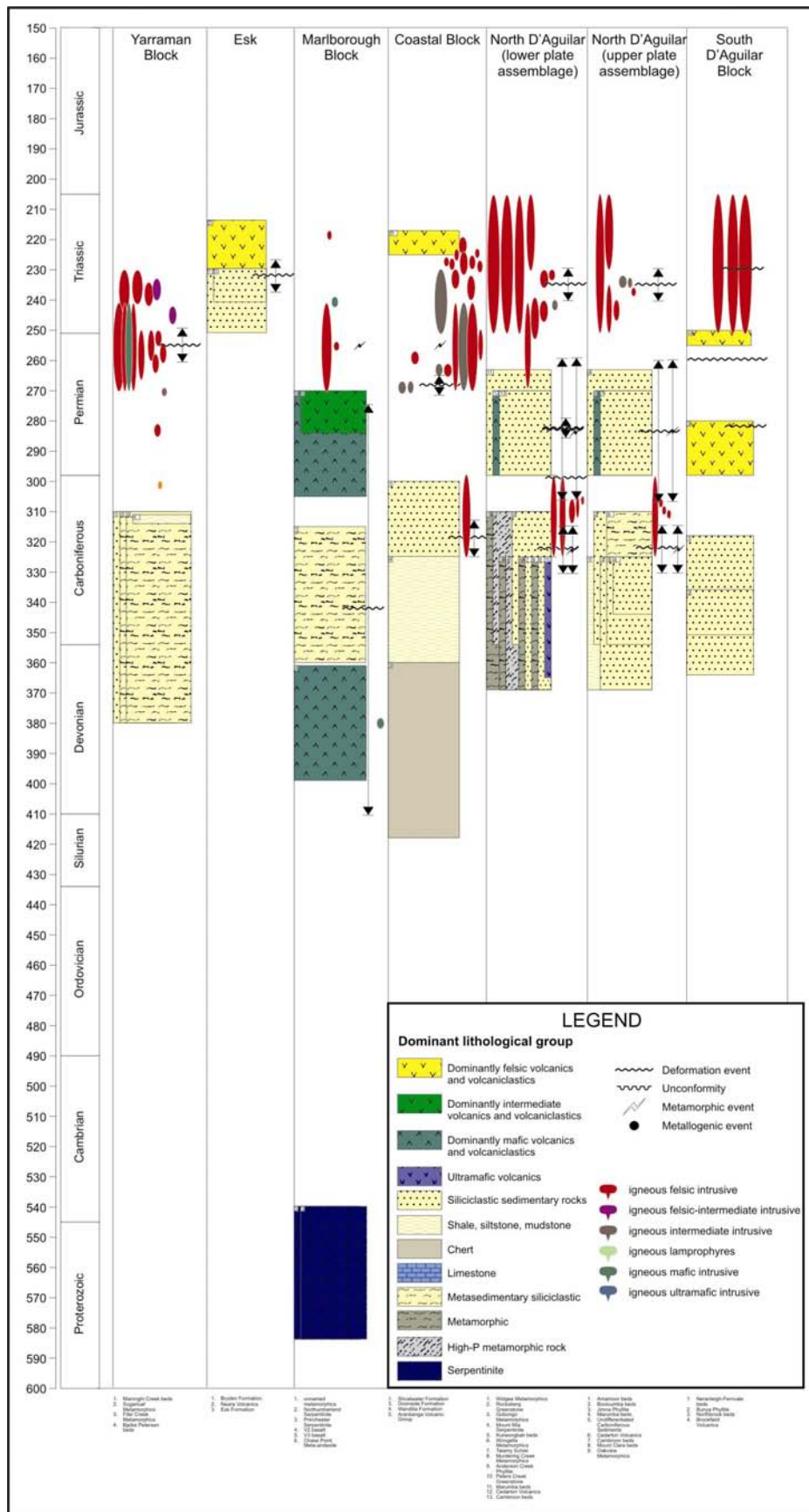


Figure 17 continued. Late Neoproterozoic to Jurassic time-space plot for part of the northern New England Orogen. Refer to text for data sources and discussion. Terranes are as outlined in [Figs 15](#) and [16](#).

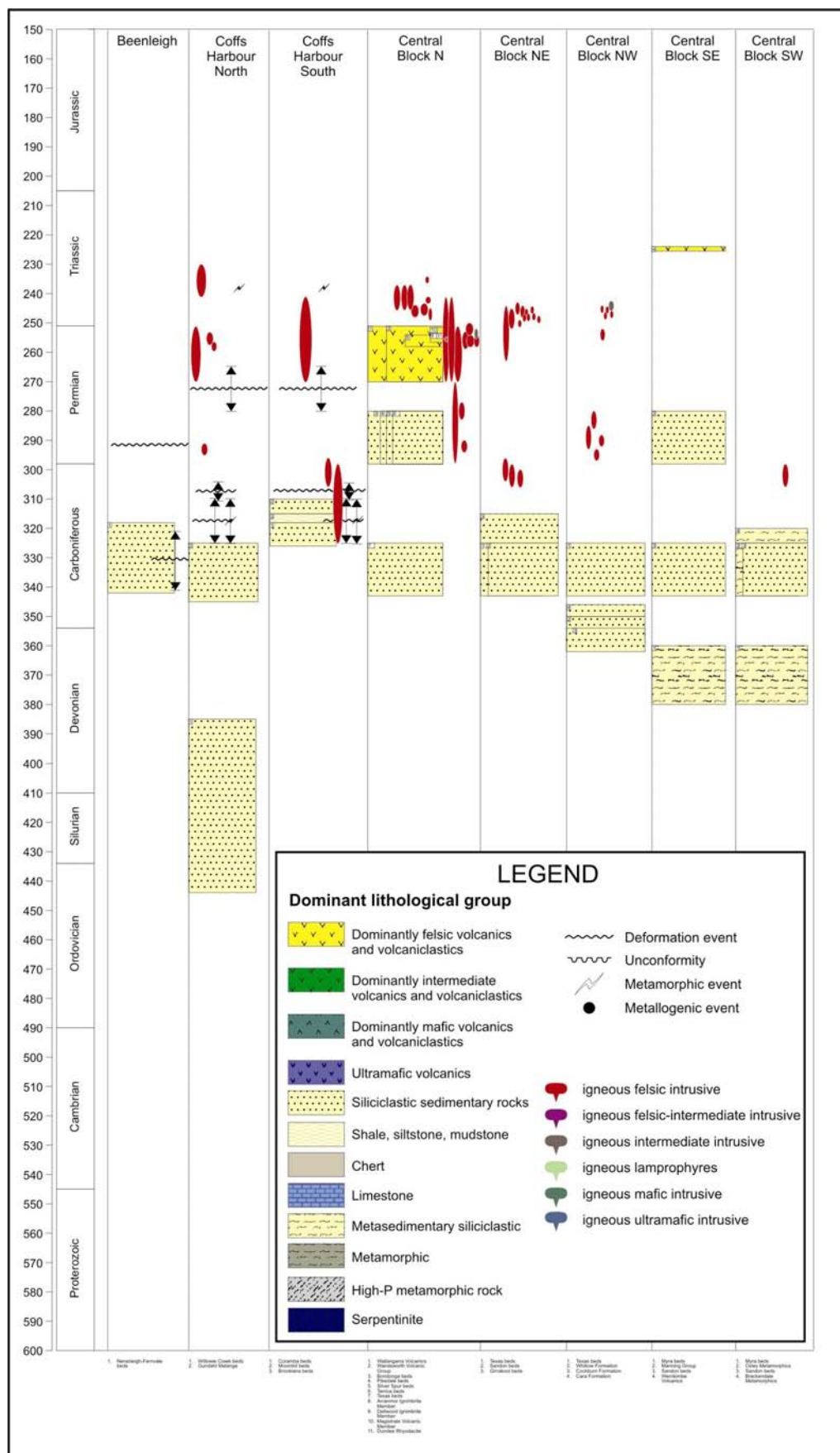


Figure 17 continued. Late Neoproterozoic to Jurassic time-space plot for part of the southern New England Orogen. Refer to text for data sources and discussion. Terranes are as outlined in Figs 15 and 16.

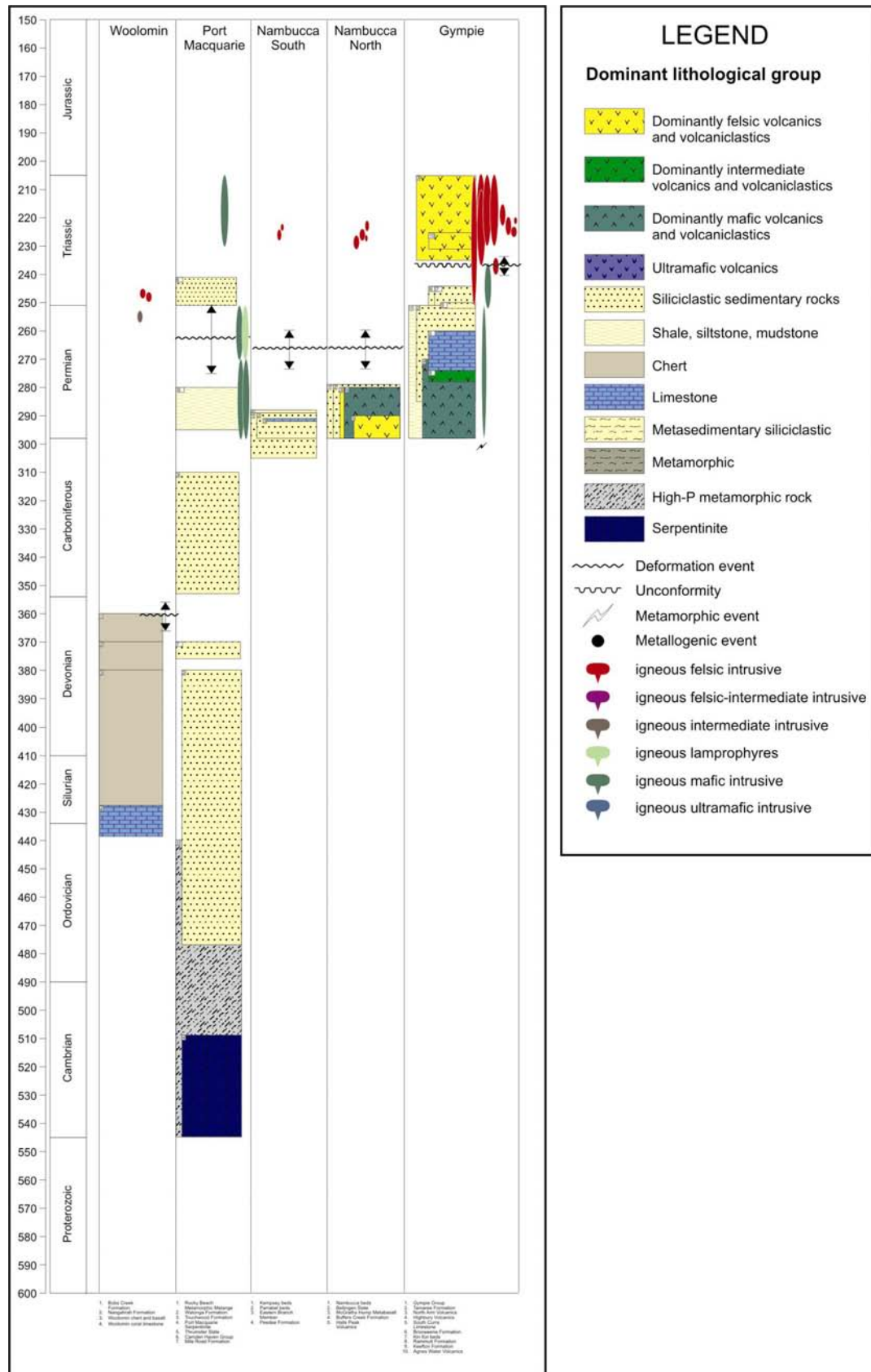


Figure 17 continued. Late Neoproterozoic to Jurassic time-space plot for part of the southern New England Orogen. Refer to text for data sources and discussion. Terranes are as outlined in Figs 15 and 16.

1.3.3. Silurian to Middle to early Late Devonian

Post-Benambran to Tabberabberan Orogeny: ca. 440-380 Ma

Late Silurian to Middle Devonian arc successions in the Gamilaroi Terrane, Calliope Volcanic Arc, Coffs Harbour North (Willowie Creek Group), Silverwood (Silverwood Group), and at Alice Creek (Craigilee Equivalents), are interpreted to have all formed as part of one intraoceanic island arc (e.g., van Noord, 1999; Figs 17, 18, 19). Island arc magmatism is also recorded along the PMFS (Early Silurian) and in the Marlborough Block (Mid Devonian). Ocean basin cherts and basalts of the Woolomin Terrane and Coastal Block were deposited in this time interval.

Late Silurian to Middle Devonian island arc – Gamilaroi-Calliope

Southern NEO

- The Gamilaroi Terrane is made up of Late Silurian-Devonian arc volcanics and volcanogenic sediments (Flood and Aitchison, 1988, 1992; Aitchison and Flood, 1995; Stratford and Aitchison, 1995; Offler and Gamble, 2002; Fig. 19). Diverse suites of volcanoclastic sediments are intercalated with regionally extensive extrusive and subvolcanic metafelsic igneous rocks of low K and calcalkaline affinity (Cawood and Flood, 1989; Aitchison and Flood, 1995; Offler and Gamble, 2002).
- Geochemical studies suggest that the Late Silurian-Early Devonian successions of the Gamilaroi Terrane formed in an intraoceanic arc setting that was rifted in the Middle to Late Devonian (Offler and Gamble, 2002).
- The sedimentary and volcanic rocks of the Gamilaroi Terrane are unconformably overlain to the south and west by Late Devonian-Carboniferous rocks of the Tamworth Belt (Aitchison and Flood, 1992). The Gamilaroi Terrane had been accreted to the Gondwana margin by the Late Devonian and is constrained by the first appearance of distinctive, westerly-derived (Lachlan Orogen) quartzite clasts in the uppermost Devonian Keepit Conglomerate (Flood and Aitchison, 1992).
- Two rock successions of Late Silurian-Middle Devonian age, the Silverwood Group and Willowie Creek beds, have been interpreted as either a southern continuation of the Calliope Arc (Day et al., 1978) or as displaced fragments of the Tamworth Group (Cawood and Leitch, 1985). We support an island arc rather than a continental margin forearc basin setting for these successions (Figs 18, 19).
- Remnant blocks recording Early Silurian island arc magmatism and metamorphism occur along the PMFS. This is recorded by a ca. 425 Ma age for hornblende cumulates and diorites from Glenrock Station (Sano et al., 2004).
- Late Silurian to Devonian sedimentation recorded in the Woolomin Terrane consists of 428-380 Ma chert and basalt deposited in an ocean basin.

Northern NEO

- The Calliope Arc consists of Late Silurian to Middle Devonian shallow marine volcanoclastic sediments with varying amounts of calcalkaline felsic to mafic volcanic rocks (Fig. 19). Geochemical studies suggest formation in either a primitive continental or island arc setting (Morand, 1993a; Offler and Gamble, 2002; Murray and Blake, 2005). Murray and Blake (2005) suggest that the data support an origin for the Silurian to Middle Devonian assemblages as exotic oceanic terranes mainly from island arcs, but may include some backarc basin rocks (see Bryan et al., 2004). The Mount Morgan Tonalite intruded the succession at 381 ± 5 Ma, and has an island arc or rifted-arc geochemistry (Golding et al., 1994; Murray, 2003). The Calliope assemblage hosts the Mount Morgan gold-copper deposit (~380 Ma; Yarrol Project Team, 1997, 2003).
- The Calliope Arc rocks are overlain by Late Devonian and younger forearc basin strata of the Yarrol Belt (see below). If exotic, the Calliope must have reached its present position by the end of the Middle Devonian; the accretion event being represented by an early Late Devonian

unconformity with the overlying Late Devonian Yarrol Belt (e.g., Korsch et al., 1990). Morand (1993b) suggested the unconformity did not relate to major orogenesis, and suggested it was more consistent with a continental arc origin for the Calliope Arc.

- Ocean basin cherts were deposited in the Coastal Block during this period, with Doonside Formation sedimentation occurring between ~418 Ma and ~360 Ma.
- Calcalkaline basalts, dolerites and gabbros and fault-bounded blocks in the Neoproterozoic Marlborough ophiolite (Marlborough Terrane) have a Middle Devonian Sm-Nd isochron age (380 ± 19 Ma) and trace element data suggestive of an intra-oceanic island arc (Bruce and Niu, 2000). These relations suggest that the arc was probably built on a Neoproterozoic oceanic crust, lay only a short distance offshore, and was accreted to Gondwana by the Early Permian (Bruce and Niu, 2000; Murray, 2007).

1.3.4. Late Middle Devonian to Late Triassic

Post- Tabberabberan Orogeny to end of the Hunter-Bowen Orogeny:

The accretion of the Gamilaroi-Calliope island arc to the Gondwana margin heralded the initiation of the New England Orogen as an Andean-style continental margin. The subsequent history of the NEO records an event history distinct from the remainder of the Tasmanides (e.g., Glen, 2005). This is not to suggest that Kanimblan- or Alice Springs-related deformations are not present, just that they do not appear to be significant in regards to the formation of the Orogen. The main phases in the NEO are:

- Development of the NEO as a continental margin magmatic arc (Connors-Auburn Arc) with associated fore-arc and accretionary wedge development, from the Late Middle Devonian to the Late Carboniferous;
- An extensional, back-arc phase from the Late Carboniferous to late Early Permian, possibly transitional from the last phase. Extension was accompanied by widespread magmatism and sedimentation, including initiation of the Sydney-Gunnedah-Bowen basin system. Extension was terminated by deformation and formation of the Texas and Coffs Harbour Oroclines;
- Renewed continental margin arc magmatism and sedimentation, with associated contraction and formation of a fold-thrust belt and retroforeland basin. This was related to the main phase of the Hunter Bowen Orogeny, occurring from late Early Permian to Middle Triassic. This orogeny effectively cratonised eastern Australia. It was followed by widespread, back-arc(?) extensional-related magmatism and sedimentation.

1.3.5. Late Middle Devonian to Late Carboniferous

Post- Tabberabberan to end of Connors-Auburn Arc: ca. 380 - 305 Ma

During the Late Devonian and Carboniferous, eastern Australia was located on a convergent margin with a westerly dipping subduction zone responsible for the development of a magmatic arc in the west, flanked by a forearc basin and accretionary wedge in the east (Cawood, 1982; Murray et al., 1987). Components of the convergent plate margin system in the southern NEO include the accretionary wedge (Woolomin, Central and Coffs Harbour blocks), forearc basin (Tamworth Belt and Hastings Block) and magmatic arc (now either buried or removed; Figs 16, 19). In the northern NEO, a continental margin magmatic arc is represented by the Connors and Auburn Arches, the forearc basin is represented by the Yarrol Belt, the accretionary complex represented by the Coastal, Yarraman, North and South D'Aguilar and Beenleigh blocks (Figs 16, 19). A possible back-arc basin occurs in the north (Drummond Basin), but not in the south (Glen, 2005), although the Late Devonian of the Lachlan Orogen is probably the backarc equivalent.

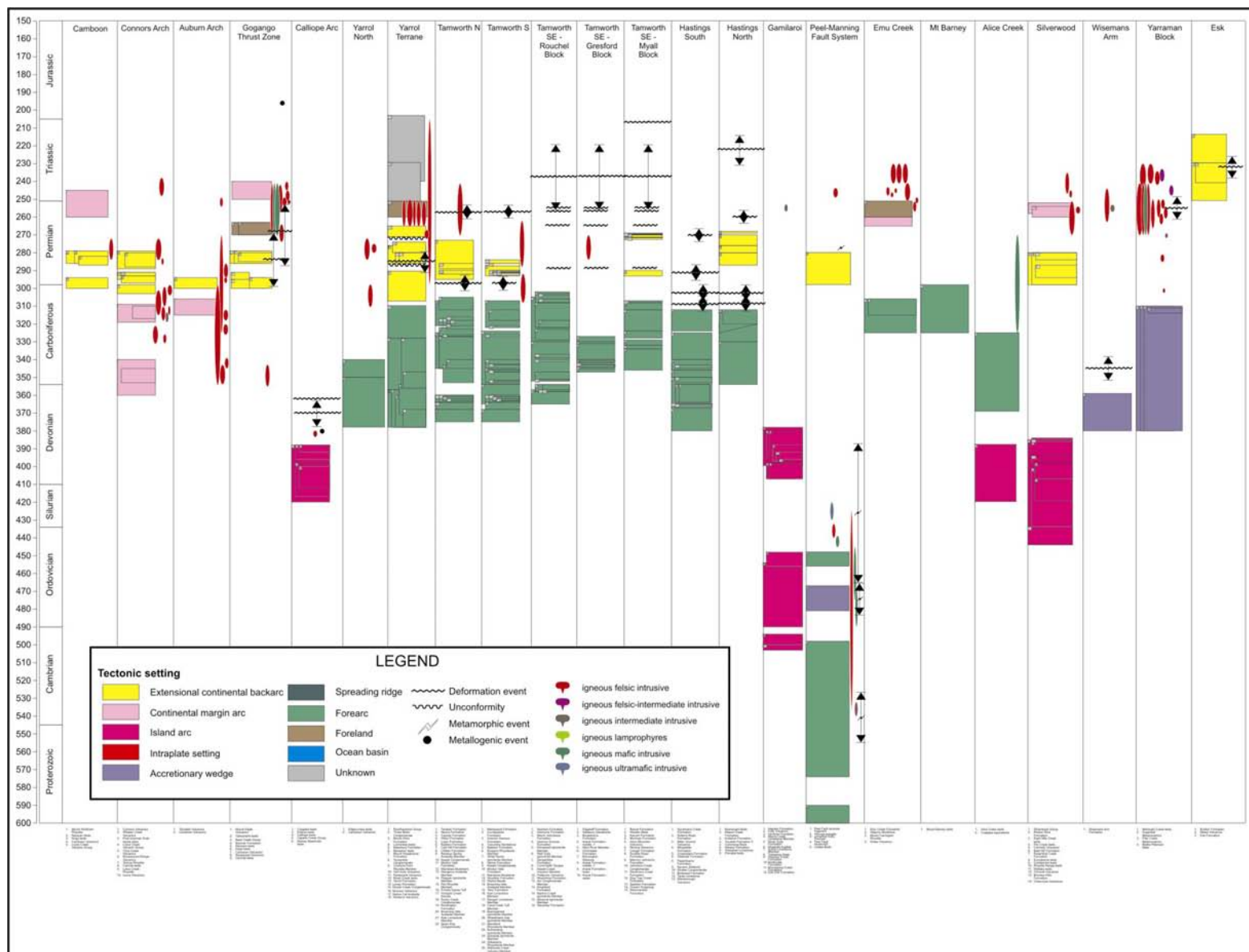


Figure 18. Time-space plot of the NEO by tectonic setting. Refer to text for data sources and discussion. Terranes are as outlined in Figs 15 and 16.

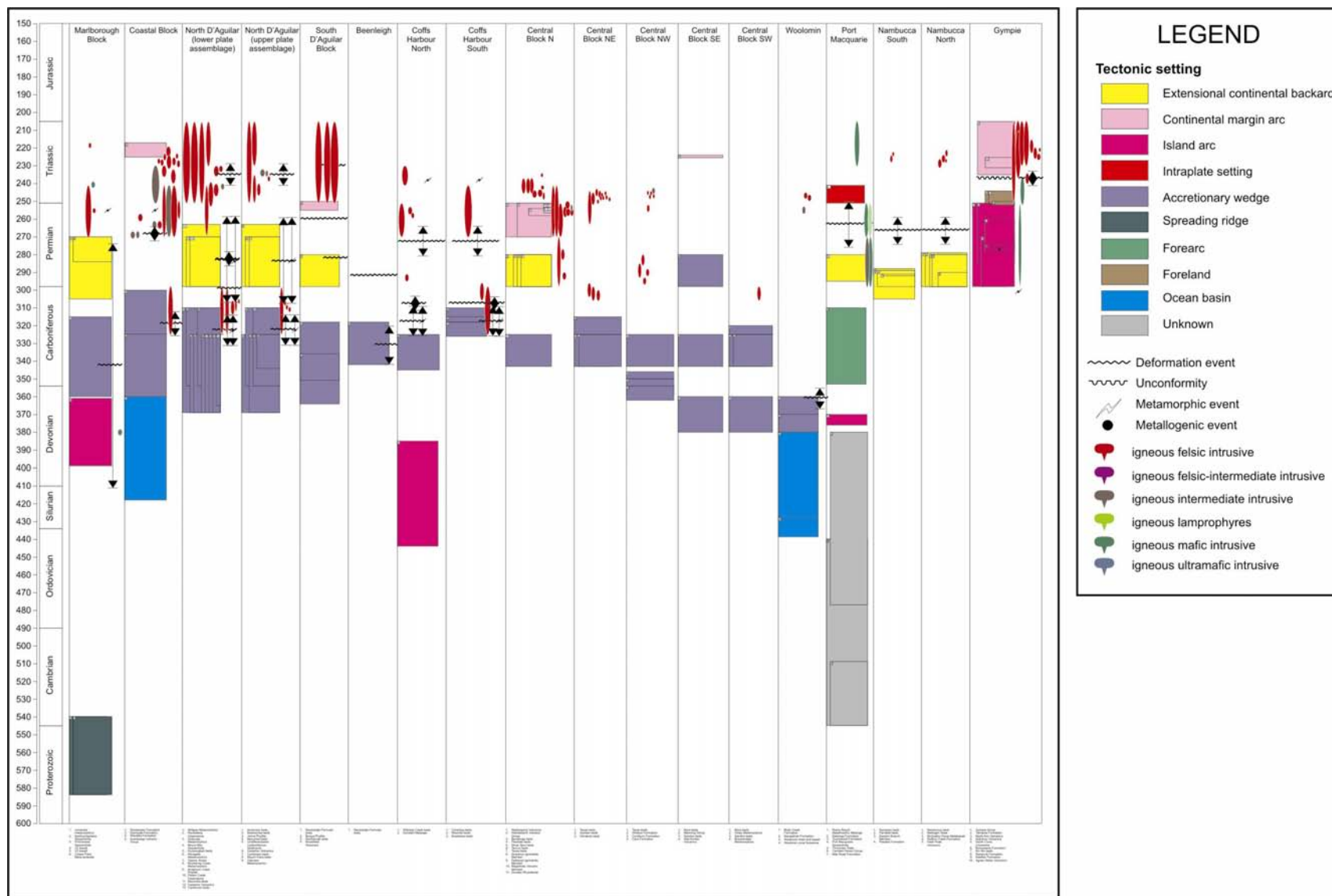


Figure 18 continued. Time-space plot of the NEO by tectonic setting. Refer to text for data sources and discussion. Terranes are as outlined in [Figs 15](#) and [16](#).

Devonian-Carboniferous continental margin magmatic arc – Connors-Auburn

- There is general agreement that, in the Late Devonian, the tectonic setting changed as a result of the (Devonian) accretion of the intraoceanic island arc (Gamilaroi-Calliope) to the Gondwanan continental margin (e.g., Aitchison et al., 1992b; although cf. Morand (1993a) for an alternate view), leading to the development of a continental margin calcalkaline arc that remained active until almost the end of the Carboniferous (Roberts et al., 1995, see below).
- The Connors and Auburn Arches consist of late Paleozoic granites and silicic volcanic rocks that are considered to represent a Late Devonian to Early Carboniferous Andean-style volcanic arc west of the Yarrol Terrane (Murray, 1986; Fig. 19); though were interpreted as back-arc by Bryan et al. (2001). The Connors and Auburn Arches are separated by the Gogango Thrust Zone which represents part of the Permian Bowen Basin succession strongly deformed and thrust westwards in the Late Permian to Early Triassic (Fergusson, 1991). Igneous activity in both arches commenced at c. 350 Ma, although the main pulse of granite formation was between c. 324 Ma and 313 Ma in the Auburn Arch and c. 316-305 Ma in the Connors Arch (Murray, 2003). Murray (2003) suggested that the older granites in the Auburn Arch and the older part of the Connors Arch are subduction-related, and the 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, at around 305 Ma (see below).

Devonian-Carboniferous forearc basin

To the west of the Peel and Yarrol Faults, the Tamworth Belt in the southern NEO and the Yarrol Belt in the northern NEO are usually interpreted to be parts of a continuous forearc basin (Korsch et al., 1990; Figs 16, 19). The forearc basin deepened towards the east and consists predominantly of continental to shallow marine clastic sediments, derived predominantly from the volcanic arc to the west.

Southern NEO

- Early and Middle Devonian volcanoclastic sediments, volcanics and limestones of the Tamworth Belt (Fig. 17) were deposited in a forearc basin to the west of the PMFS. Cawood and Leitch (1985) interpreted the Tamworth Group as the fill of a forearc basin between the western arc and the partly uplifted accretionary prism to the east of the PMFS. An arc along the inboard part of the orogen is not preserved but is inferred from large Late Devonian olistromal blocks of andesitic volcanic rocks in the inboard part of the forearc basin, the Tamworth Belt (Brown, 1987). Indeed, strong evidence for its existence is provided by the composition of sandstones from the Tamworth Belt and accretionary complex with virtually all of them being arc-derived volcanoclastics (Cawood, 1983; Korsch, 1984), whereas volcanic material is more abundant closer to the active magmatic arc (McPhie, 1987). The forearc Tamworth Belt was filled by marine strata that become finer grained and of deeper water character eastwards towards the PMFS, which marks the preserved outboard edge of the basin (Yarrol Project Team, 1997). While deposition in the forearc basin was dominated by volcanoclastic sediments with minor interbedded volcanics, limestones developed in the Early Carboniferous during periods of diminished terrigenous sedimentation (Roberts and Engel, 1980; Cawood and Leitch, 1985). More detailed descriptions of sedimentary rocks from the Tamworth Belt are given by Cawood (1983) and Korsch (1984).
- The Hastings Terrane is widely considered to represent a dispersed fragment of the Tamworth forearc basin, with which it shows an overall similarity throughout its tectonic development (Cawood and Leitch, 1985; Roberts et al., 1993). Units interpreted to part of the Tamworth Belt form part of the Texas and Coffs Harbour Oroclines. Components of the Oroclines considered to be part of the forearc basin succession include rocks of the Emu Creek Block and Carboniferous outcrops at Mount Barney and Alice Creek (Figs 17, 18, 19). We have also assigned a forearc environment for deposition of the Touchwood Formation in the Port Macquarie Block.
- Volcanism and deposition in the Tamworth Belt appear to have largely ceased at ~305 Ma in response to the eastward migration of the subduction zone (Roberts et al., 2004). Mechanisms to account for this migration have included slab breakoff (Caprarelli and Leitch, 2001) and rollback (Jenkins et al., 2002).

Northern NEO

- The Rockhampton Group in the Yarrol Terrane is considered to have been deposited in a forearc basin (Murray et al., 2003; although cf. Bryan et al., 2001) and covered the former Late Devonian arc, as well as older basin strata to the east (Fig. 19). Late Devonian strata consist of volcanoclastic sandstone and conglomerate, derived from the andesitic arc to the west, interbedded with sediments and some limestone (Yarrol Project Team, 1997).
- On a regional scale, the Late Devonian rocks of the Yarrol Terrane are considered to represent a transitional phase in the change from an intraoceanic setting (Calliope Arc), epitomised by the Middle Devonian Capella Creek Group, to a continental margin setting in the northern NEO in the Carboniferous (Murray and Blake, 2005).

Devonian-Carboniferous accretionary wedge

The Woolomin, Wisemans Arm, Central and Coffs Harbour blocks in the southern NEO and the Coastal, Yarraman, North D'Aguilar, South D'Aguilar and Beenleigh blocks in the northern NEO are interpreted as a once continuous accretionary wedge that grew oceanwards by accreting trench-fill volcanoclastic turbidites (derived from a magmatic arc) and minor amounts of oceanic crust (basalt, chert, mudstone) (Figs 16, 19; Korsch et al., 1990).

Southern NEO

- The mid-Devonian to Carboniferous accretionary wedge in New South Wales (Fig. 19) is made up of the Woolomin Terrane (chert, basalt, mostly accreted ocean floor, very rare volcanoclastic sandstone and mudstone), the Central Block (N, NE, NW, SE, SW, turbidite, chert, basalt – mixed ocean floor and trench fill), the Coffs Harbour Block (mainly trench fill turbidite), southern Beenleigh Block and successions at Wisemans Arm.
- The accretionary wedge consists largely of deep marine sedimentary rocks, including metabasalt-chert-argillite associations (Cawood, 1982). 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 of the westernmost Woolomin Terrane (Aitchison et al., 1992b). The Central Block(s) consist of deformed and dismembered volcanogenic siltstone and sandstone, basalt, chert and minor conglomerate (Cross et al., 1987).

Northern NEO

- Mid-Devonian to Carboniferous accretionary wedge rocks in the northern NEO (Fig. 19) occur in the North and South D'Aguilar, Yarraman and Beenleigh blocks in the south and the Coastal Block in the north. The Beenleigh Block consists of greywacke, argillite and Early Carboniferous chert (Aitchison, 1988). Similar strata occur in the South D'Aguilar Block (Holcombe et al., 1997b). In contrast, the North D'Aguilar Terrane is a composite terrane, containing high-level accretionary wedge rocks as well as ophiolitic components of an accretionary wedge that were subducted to more than 18km before 315 Ma, and were subsequently exhumed in the lower plate below a latest Carboniferous, low angle normal fault (Little et al., 1992, 1995; Holcombe et al., 1997b). Similar-aged accretionary wedge successions are recorded in the Yarraman and Marlborough Blocks.

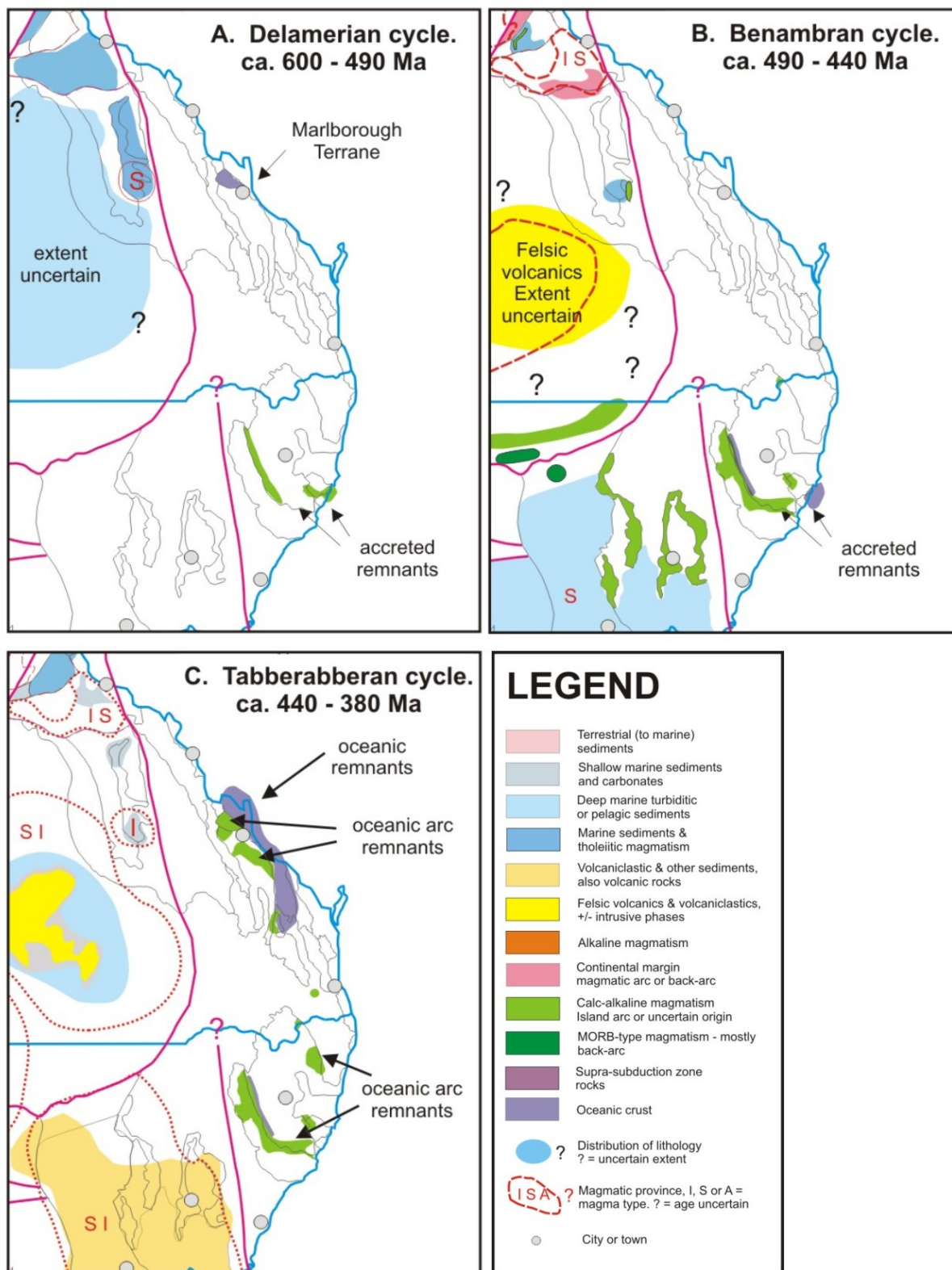


Figure 19. Generalised distribution of rocks in the New England Orogen, by tectonic cycle. A = Rodinian to Delamerian, B = Benambran, C = Tabberabberan.

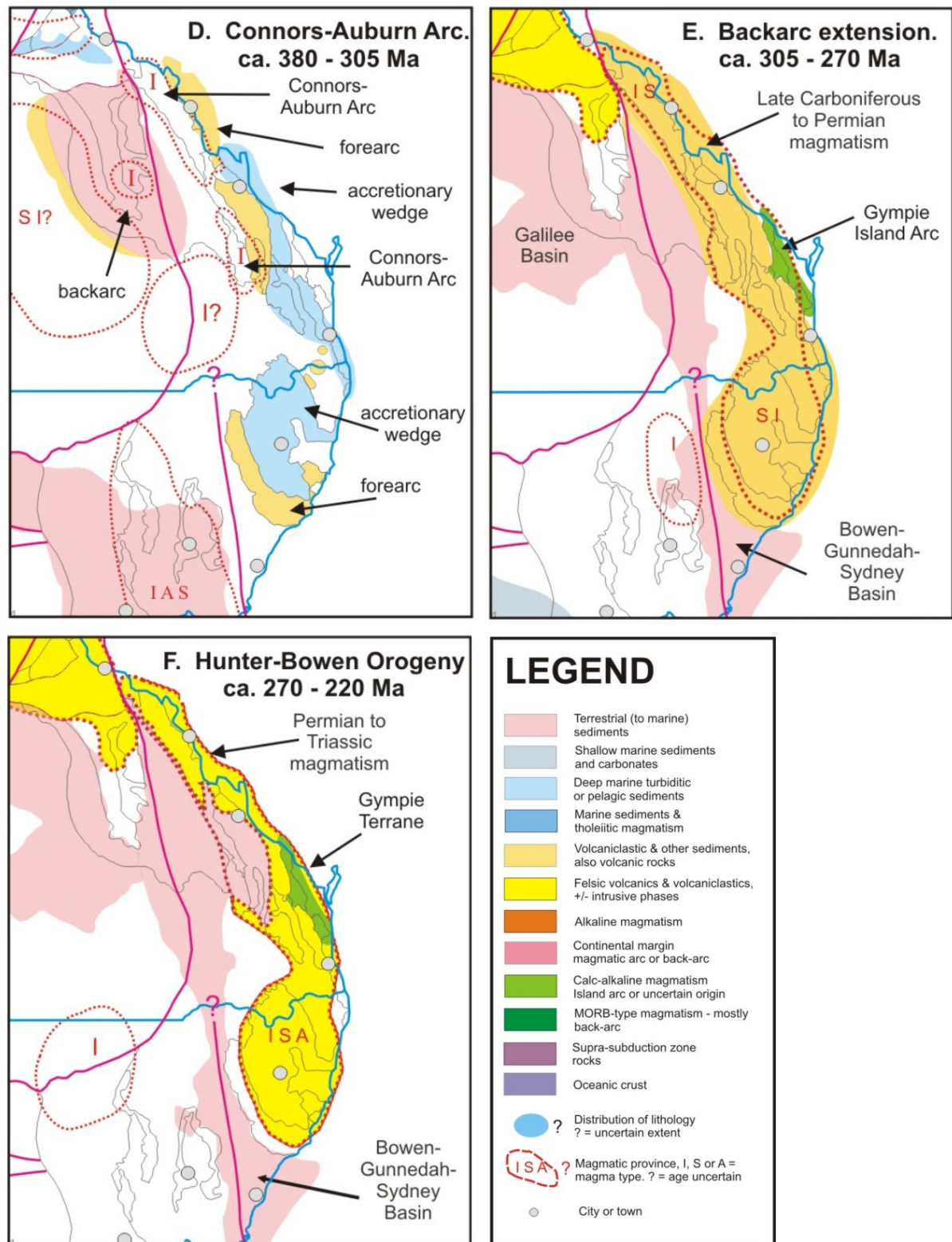


Figure 19 continued. Generalised distribution of rocks in the New England Orogen, by tectonic cycle. D = Connors-Auburn Arc (~Kanimblan and younger), E = backarc extension phase (younger part of post-Kanimblan cycle, F = Hunter-Bowen.

1.3.6. Late Carboniferous to late Early Permian

Back-arc extension to orocline formation: ca. 305 to 265 Ma

Along the Gondwana margin, a transition took place from active accretion in the mid-Carboniferous to widespread extension through the Late Carboniferous-Early Permian (Leitch, 1988; Holcombe et al., 1997a). This transition has been interpreted in terms of eastward retreat of the subducting slab, and migration of the volcanic arc offshore (Holcombe et al., 1997a). Thus, by the Early Permian, much of the NEO was the site of numerous backarc extensional basins filled with marine and terrestrial deposits and mafic-silicic volcanics. Two recent models for the Late Palaeozoic evolution of the southern NEO (Caprarelli and Leitch, 1998, 2001; Jenkins et al., 2002) involve alternate mechanisms, namely slab breakoff, and arc rollback, respectively, to explain the timing of easterly migration of the subduction zone, generation of granitic suites, large scale volcanism and deformation.

In the Early Permian (~295-280 Ma), the initiation of the Bowen-Gunnedah-Sydney basin system occurred to the west of the continental margin magmatic arc in a backarc tectonic setting (Korsch et al., 1993; Fig. 19). Also around this time, but probably after extension ceased, oroclinal bending of the forearc and accretionary wedge successions (~285-265 Ma) produced the Texas and Coffs Harbour Oroclines (Korsch and Harrington, 1987; Murray et al., 1987).

- In the southern NEO, the cessation of subduction just before the end of the Carboniferous and a postulated eastward migration of the subduction zone was followed by emplacement of early S-type granites at ~300 Ma (e.g., Collins et al., 1993; Fig. 19). Approximately contemporaneous cessation of deposition in the Tamworth forearc basin at c. 305 Ma was followed by major strike-slip faulting and anticlockwise rotation of crustal blocks (Roberts et al., 1995, 1996). A change from contraction to extension in the NEO led to the region being located in a backarc environment (Roberts et al., 1996).
- From the latest Carboniferous through the Early Permian, deposition of bimodal volcanics, volcanoclastic and siliciclastic sedimentary rocks in an extensional continental backarc setting occurred throughout most of the NEO (e.g., Camboon, Connors Arch, Auburn Arch, Gogango Thrust Zone, Yarrol, Tamworth Belt, Hastings Block, Peel-Manning FS, Silverwood, Marlborough Block, Central Block N, North D'Aguilar, South D'Aguilar, Nambucca Block and Port Macquarie; Figs 17, 18, 19). These rocks overlie the older forearc and accretionary wedge successions.
- Latest Carboniferous to Early Permian extension is also recorded by the emplacement of granites into, and formation of low-angle extensional faults in, the former accretionary wedge (Glen, 2005). Late Carboniferous granitoids of the New England Batholith primarily intruded into the accretionary wedge assemblage, with minor intrusion into the Tamworth Belt. This eastward shift in magmatism, at ca. 300 Ma, into the former accretionary wedge, is interpreted to reflect the initial outboard retreat of the arc in the southern NEO (Collins et al., 1993). This magmatism is represented by the S-type Hillgrove and Bundarra Granite Suites (Collins et al., 1993; Kent, 1994). The S-type granitoid suites were emplaced before tectonism associated with the contractional Late Permian Hunter-Bowen Orogeny (see below), whereas I-type suites intruded in the Late Permian-Triassic (Shaw and Flood, 1981).
- Both Caprarelli and Leitch (1998, 2001) and Jenkins et al. (2002) proposed that Late Carboniferous volcanism ceased at around 305 Ma, prior to the commencement of either slab breakoff or rollback. Intrusion of the Hillgrove Suite at 302 Ma (Collins et al., 1993; Kent, 1994) was interpreted by Caprarelli and Leitch (1998) as a response to the melting of sediments by upwelling of asthenosphere, following slab breakoff, with intrusion occurring in a contractional environment. Jenkins et al. (2002), however, suggested that rollback by 300 Ma had caused magmatism to move eastward into the accretionary wedge, resulting in generation of the Hillgrove Suite in an extensional environment. We prefer the latter interpretation.
- The Nambucca rift basin formed during backarc extension in the Early Permian. Sediments in the Nambucca Basin were deposited unconformably onto the forearc successions of the Hastings

Block (Roberts et al., 1993), the implication being that the Hastings Block was in its current position prior to Early Permian sedimentation in the Nambucca Block (Johnston et al., 2002). This, however, may not necessarily be the case; the Permian rocks of the Nambucca Block are highly deformed and they may have been caught between the Coffs Harbour Block moving south and the Hastings Block moving north during the period of oroclinal bending (R.J. Korsch, pers. comm. 2008).

- Tholeiitic and alkaline dolerite dykes with enriched geochemical signatures intruded the ophiolite succession of the Marlborough Terrane in the Early Permian (293±35 Ma, Bruce and Niu, 2000). This magmatism is characterised by compositions typical of intraplate basalts, and a continental intraplate origin is preferred by Bruce and Niu (2000).
- Extensional events in the Early Permian caused subsidence of the Sydney-Gunnedah Basin, abundant basaltic and rhyolitic volcanism and onlap of the forearc basin (Leitch, 1988; Scheibner and Basden, 1998; Roberts and Geeve, 1999) by the Permian sediments (Roberts et al., 2004).
- The Camboon Volcanics are widely considered to represent a superimposed, Early Permian subduction-related episode, although associated forearc basin or accretionary wedge elements have not been recognised (Holcombe et al., 1997a; Withnall et al., 1998). The main units identified in the Camboon Province are the Camboon Volcanics, Lizzie Creek Volcanics and the Nogo and Narayen beds (Figures 17 and 18), which we have interpreted as an extensional continental backarc setting. Holcombe et al. (1997a) also have suggested that the Camboon Province represents an extensional event that is not necessarily related in any aspect to active subduction.
- The time of oroclinal bending that produced the Texas-Coffs Harbour Oroclines has been the subject of much debate. Some authors consider that it occurred in the Late Carboniferous (310-300 Ma, Murray et al., 1987), Early Permian (290-280 Ma; Fergusson and Leitch 1993), and Early to mid-Permian (280-265 Ma; Korsch and Harrington, 1987). Offler and Foster (2008) suggest that development of the Texas and Coffs Harbour Oroclines took place between 273-260 Ma. We consider oroclinal bending most likely took place after Early Permian backarc extension but prior to the main phase of the Hunter-Bowen Orogeny.

Gympie Island Arc

- An inferred Early Permian backarc setting for the NEO and Sydney-Bowen Basin requires that an arc environment existed outboard at this time (Fig. 19). The Early Permian Highbury Volcanics of the Gympie Terrane have a chemical signature indicative of an arc setting (Sivell and McCulloch, 1997). The submarine and subaerial island arc tholeiites, basaltic tuff breccias and lavas that constitute the Highbury Volcanics are unconformably overlain by andesites and dacites of the Rammutt Formation (Sivell and McCulloch, 2001). The primitive chemical and isotopic character of these units implies a juvenile island-arc terrain, isolated from the influence of continental crust (Sivell and Waterhouse, 1988). This is consistent with their intimate association with primitive oceanic backarc (Cedarton and Cambroon) basalts (Sivell and McCulloch, 1997), which suggests that a well-developed backarc separated the primitive Gympie island arc from the NEO in the Early Permian.
- The intraoceanic Gympie island arc was located east of the continental margin of Gondwana in the Early Permian. Detrital zircon data from Gympie (Korsch et al., in press c) suggest that the Gympie arc was attached back to eastern Australia at the end of the Permian or start of the Triassic.

1.3.7. Late Early Permian to Middle Triassic

The Hunter-Bowen Orogeny: ca. 265 Ma to ca. 230 Ma

The late Early to early Late Permian (~265-262 Ma) saw another change in the dynamics of the subduction system, when a continental margin magmatic arc was re-established along the Palaeo-Pacific continental margin of Australia and the backarc changed from an extensional to a contractional regime (Korsch and Totterdell, 1995). This led to the formation of a retroforeland fold-thrust belt west of the magmatic arc, which was better developed in the northern NEO than the southern NEO (Korsch et al., 1997). This contractional regime resulted in the development of a major retroforeland basin phase in the Bowen-Gunnedah-Sydney basin system that continued until the Middle Triassic (Korsch and Totterdell, 1995).

Thus, this period is characterised by renewed arc magmatism, the foreland basin stage of development of the Bowen-Gunnedah-Sydney basin system and the Hunter-Bowen Orogeny (Fig. 19).

The Late Triassic (c. 230 Ma) saw a switch in geodynamics back to an extensional, probably backarc environment. This resulted in a change in plutonism to A-type granites, bimodal volcanism and development of extensional basins with coal-bearing successions. This also marked the timing of effective cratonisation of eastern Australia.

- A change from extensional to contractional tectonism began in the latest Early Permian, at ca. 270 Ma (Roberts et al., 1996). The Bowen-Gunnedah-Sydney basin system developed as a foreland basin phase in a backarc setting that persisted until the Middle Triassic (Korsch et al., in press a). Foreland basin sedimentation is also recorded in the Gogango Thrust Zone, Yarrol Terrane, Emu Creek Block, and in the Gympie Terrane (Figs 17, 18, 19).
- The late Early to early Late Permian saw the renewed onset of subduction-related magmatism with voluminous Late Permian to Early (and Middle) Triassic intrusive and extrusive magmatism occurring throughout the NEO (Gust et al., 1993; Holcombe et al., 1997b; Van Noord, 1999; Fig. 19). Examples of arc-related volcanism include Late Permian to Early Triassic deposition of Mount Wickham Rhyolite (Camboon Province) and Mount Eagle Volcanics (Gogango Thrust Zone) in a continental margin magmatic arc setting (Fig. 18). Continental margin arc-related rocks also occur at Emu Creek, Silverwood, and Central Block N, with Triassic continental margin arc volcanic rocks in the Gympie Terrane (Fig. 18). Volcanism is predominantly andesitic (Holcombe et al., 1997b).
- Widespread intrusive magmatism also occurred at this time. These granites extend from the New England Batholith (Shaw and Flood, 1981), in the south, at least up to Rockhampton in the north (e.g., Murray, 2003; Fig. 19). About half of the exposed granitoids in the northern NEO have K-Ar ages between 270 Ma and 230 Ma (Gust et al., 1993; Murray, 2003), although most of these are reset ages. The granitoids are widely distributed and are predominantly of intermediate to felsic, I-type, composition, and most seem to belong to the Clarence River Supersuite or equivalents (Bryant et al., 1997; Murray, 2003). They are compositionally very similar to the earlier (Late Carboniferous to Early Permian) magmatism (Murray, 2003). Some A-types are also apparently present (Murray, 2003).
- The widespread calcalkaline magmatism strongly supports active subduction during this time with compositional changes probably related to changes in arc configuration. Gust et al. (1993), for example, suggested it may reflect changes in the angle of the subducting slab. As documented by Gust et al. (1993), magmatism appears to wane in the Middle Triassic, although age control is poor.
- This time interval encompasses the Mid-Permian to Middle Triassic orogenic event (which includes deformation, metamorphism, magmatism, foreland sedimentation) known as the Hunter-Bowen Orogeny (Murray, 1997a, b; Holcombe et al., 1997b; Roberts et al., 2006; Korsch et al., in press b). In the New England Orogen, this phase of deformation is characterised by retrothrusting driven by subduction further to the east. Subsidence of the Bowen and Gunnedah basins during the

foreland basin phase was driven by thrust loading due to westward-propagating thrust sheets from the New England Orogen (Korsch et al., in press b). The orogeny covers a period of about 35 m.y. from ~265 Ma to ~230 Ma, (Holcombe et al., 1997b; Korsch et al., in press b). The foreland basin phase of sedimentation associated with the Hunter-Bowen Orogeny was punctuated by a series of discrete contractional events (Figure 17), frequently producing unconformities which were very short-lived (Korsch et al., in press b). Contractional events of the Hunter-Bowen Orogeny thrust the Tamworth Belt westwards over the eastern edge of the Sydney-Gunnedah Basin and Lachlan Craton (Korsch et al., 1997; Roberts et al., 2004) in the Late Permian and earliest Triassic.

- The Gympie region appears to have been the site of a continent-island arc collision and recent detrital zircon age spectra from the Rammutt and Keefton Formations (Korsch et al., in press c) provide some constraints on the accretion of the island arc component of the Gympie Terrane to the eastern Australian margin. Zircon age data of Korsch et al. (in press c) suggest that the Gympie Terrane came into contact with, and was sourced from, the accretionary wedge prior to deposition of the Keefton Formation (~250 Ma; Permo-Triassic boundary).
- In contrast to the intermediate-dominated composition of Early and Middle Triassic granitoids in the northern NEO, the Late Triassic (ca. 230-220 Ma) is characterised by intrusions of dominantly silicic granite composition associated with the development of volcanic complexes of rhyolite and minor mafic lavas (Stephens et al., 1993; Holcombe et al., 1997b), including A-type magmatism (D. Champion, pers. comm., 2008). The change from backarc contraction and thrust loading to backarc extension was probably at ~230 Ma (R.J. Korsch pers. comm. 2008).
- Permian-Triassic plutonism was followed by widespread Late Triassic extension, characterised by the development of small, elongate basins, associated with bimodal volcanics and interbedded coal-bearing successions (e.g., Ipswich, Tarong, Lorne, Clarence-Morton basins; Holcombe et al., 1997b). Plutonism waned at this stage, and was restricted to small plutons near the coast. Therefore, following the Hunter-Bowen Orogeny, it appears that the New England Orogen reverted to a retreating subduction boundary system during the Middle-Late Triassic (Jenkins et al., 2002). This includes the Middle to Late Triassic deposition of sediments of the Esk Trough, also most probably in an extensional continental backarc setting (R.J. Korsch, pers. comm., 2008).

1.4. Thomson Orogen, and cover basins, and Koonenberry Belt

by E Mathews, DC Champion, N Kositsin and C Brown

Introduction

The regions covered in this section include the Thomson Orogen – specifically the Anakie Inlier and Fork Lagoons Province (Queensland) and the Louth-Bourke region (New South Wales), the northeastern part of the Delamerian Orogen – the Koonenberry Belt (New South Wales), and various Palaeozoic basins (Adavale, Drummond, Galilee, Bowen-Gunnedah-Sydney, Surat) mostly overlying the Thomson Orogen or its contact with the New England Orogen (refer to [Figure 1b](#)). Although the Charters Towers region most probably belongs to the Thomson Orogen (e.g., Kirkegaard, 1974), it is chiefly discussed in the Northern Queensland section of this report, with which it shares many similarities. The Warburton and Cooper Basins, although not specifically dealt with in this report are included in the time-space plot for this region ([Fig. 20](#)) and in the discussion for this section.

The Thomson Orogen is perhaps the least understood of the eastern Australian orogens, largely because of the thick succession of overlying basin sediments. Similarly, apart from the southern and northern boundaries, the full extent of the Thomson Orogen is poorly defined. The northern and western boundaries of the Thomson Orogen (and the western boundary of the Lachlan Orogen) are largely defined by the enigmatic Tasman Line, which is traditionally considered to separate Proterozoic rocks from Phanerozoic eastern Australia (see Direen and Crawford, 2003). Apart from the type area in the north, however, where it is clearly evident based on geophysical anomalies that truncate the Proterozoic Mount Isa Province, for much of its length it is nebulous and difficult to identify unequivocally. This has resulted in numerous ‘Tasman Line’ interpretations in terms of both location and tectonic significance (see Direen and Crawford, 2003). For this reason, and the extensive basin cover, we have not attempted to fully delineate the western boundary of the Thomson Orogen north of the Koonenberry region ([Figs 1a, 1b, 21](#)). The eastern boundary with the New England Orogen is also poorly defined, because of overlying basin cover (i.e. Bowen Basin; [Fig. 1b](#)). We have based our eastern boundary on new Nd isotopic data (unpublished GA-GSQ data), which suggests that the boundary is east of that shown by Glen (2005). The recent Thomson-Lachlan seismic survey images the southern boundary with the Lachlan Orogen, and it is defined by the major planar, north dipping Olepoloko Fault (Glen et al., 2007c).

Little is known about the age or nature of the basement to the Orogen ([Fig. 21](#)). Most information is based on the Anakie Inlier in Queensland and also from sparse drillhole data which has intersected the Orogen (Murray, 1994), and on recent geochronology (reported in Draper, 2006). Additional new information is also becoming available with recent work, including seismic acquisition and interpretation, in the New South Wales part of the Orogen (e.g., Glen et al., 2007; Watkins, 2007). Originally, the Thomson Orogen was defined by Kirkegaard (1974) as being Cambrian to Carboniferous in age and hence equivalent to the Lachlan Orogen. New seismic data across the southern part of the Thomson Orogen show the two orogens, at least in the transect area, have different lower crustal characters, with the Thomson possessing thicker crust (Moho at 48 km) than the Lachlan (Moho at 32 km; Glen et al., 2007c). The thinner, more reflective crust of the Lachlan confirms a major difference in the crustal character between the two orogens (Glen et al., 2007c). In addition, Thomson basement rocks, at least locally (Charters Towers, Anakie Inlier; [Figs 20, 21](#)), consist of Neoproterozoic and Early Cambrian metasedimentary rocks, and contain evidence for the Delamerian Orogeny (Withnall et al., 1995; Fergusson et al., 2007c), making them different to the majority of the Lachlan Orogen. Dissimilar lithologies suggest that the depositional setting of the Thomson Orogen was not related to that of the Lachlan Orogen (e.g., Kirkegaard, 1974; Draper, 2006). Recent geochronological studies (Black, 2005, 2006, 2007; Draper, 2006) have provided important age constraints for the Thomson Orogen.

1.4.1. Late Neoproterozoic to early Cambrian

Rodinian breakup (pre-Delamerian): ca. 600 to 520 Ma

Geological and tectonic summary

Rocks of this age are best represented in the Anakie Inlier, but also occur in the Charters Towers region, Koonenberry Belt and central Thomson Orogen (Figs 20, 21). All these regions, except for the central Thomson Orogen, have good age constraints. The tectonic environment of this period is typically interpreted as passive margin and/or rifting related to Rodinian breakup of the Precambrian supercontinent (e.g., Crawford et al., 2003a; Fergusson et al., 2007c). The Late Neoproterozoic to early Cambrian (ca. 580 to 500 Ma) saw deposition of ?marine sediments, as well as lesser tholeiitic and/or alkaline magmatism, in the Charters Towers region, in the Anakie Inlier (Anakie Metamorphic Group) and in the Koonenberry Belt (Withnall et al., 1995; Hutton et al., 1997; Fergusson et al., 2001; 2007c; Gilmore et al., 2007; Figs 20, 21). Geochronology indicates that the Anakie Metamorphic Group correlates with latest Neoproterozoic-Cambrian units in the Adelaide Fold Belt (South Australia) and the Wonominta Block-Koonenberry Belt (western New South Wales) (Fergusson et al., 2001). Hence, equivalent events occurred synchronously in the Adelaide Fold Belt, Wonominta Block and Lolworth-Ravenswood Block and Charters Towers region (i.e., Cape River Metamorphics) (Withnall et al., 1995; Fergusson et al., 2001), and possibly in the central Thomson Orogen. The Anakie Metamorphic Group likely formed on a passive margin after breakup, but may also have been related to splitting or rifting of a younger microcontinent from the Gondwanan margin (e.g., Fergusson et al., 2001).

Detrital zircon ages obtained by Fergusson et al. (2007c) show that two major rock successions were sourced from the east Gondwana margin. The older succession is late Neoproterozoic in age (ca. 600 Ma) and includes the Cape River Metamorphics and lower Argentine Metamorphics (Charters Towers region) and Bathampton Metamorphics (Anakie Inlier). According to Fergusson et al. (2007c) this succession developed in a passive margin environment related to Rodinian rifting. Importantly, these successions generally contain abundant ca. 1200 Ma detrital zircons and only minor 1870-1550 Ma zircons, indicating at best only minor input from cratonic regions such as Mount Isa and Georgetown (Fergusson et al., 2007c). The younger succession is Early Palaeozoic in age and includes the Ordovician (-Silurian) Fork Lagoons beds (Anakie Inlier) and Cambrian-Ordovician upper Argentine Metamorphics (Charters Towers region). Fergusson et al. (2007c) suggested these formed in a backarc environment that developed on the former passive margin.

Geological history and Time-Space plot explanation

Anakie Inlier (Anakie region)

- The Anakie Metamorphic Group consists of widespread remnants of largely marine metasedimentary rocks (Withnall et al., 1995; Fergusson et al., 2007c; Figs 20, 21). Detrital zircon and monazite ages from the Anakie Metamorphic Group suggest a range of ages for this group. These include maximum deposition ages of 1300-1000 Ma and ca. 580 Ma for the Bathampton Metamorphics (Fergusson et al., 2001), suggesting deposition continued until the latest Neoproterozoic. Detrital zircon and monazite ages from the Wynyard Metamorphics show three age components at ca. 580-570 Ma, ca. 540 Ma, and ca. 510 Ma (Fergusson et al., 2001, 2007c). Sediments probably relate to, or were derived from, Rodinian rifting along the Gondwanan passive margin (Fergusson et al., 2001, 2007c). Detrital zircon ages of the Anakie Metamorphic Group indicate that a Grenville-aged orogenic belt may have existed in northeastern Australia to provide the major sediment source (Fergusson et al., 2001). There is little evidence for significant sourcing of sediments from older Proterozoic terranes such as Mount Isa or Georgetown (Fergusson et al., 2007c). The Anakie Inlier is thought to extend undercover to at least the Nebine Ridge to the south and it is possible that correlates of the Anakie Metamorphic Group extend to there (e.g., Withnall et al., 1995).

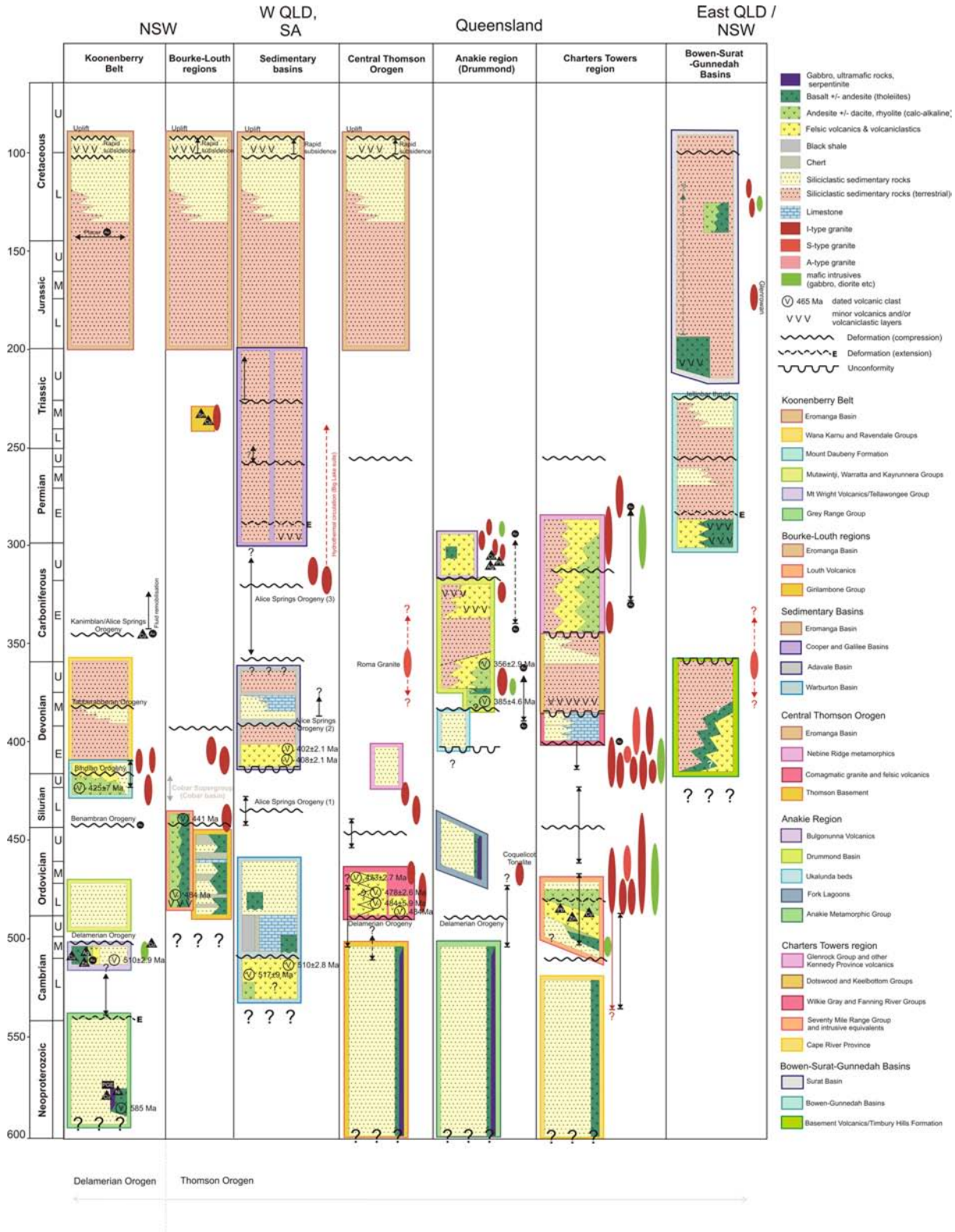


Figure 20. Late Neoproterozoic to Cretaceous time-space plot for the Thomson Orogen, Koonenberry Belt, and cover basins. Refer to text for data sources and discussion. Orogens, regions and basins are as shown in Fig. 1b.

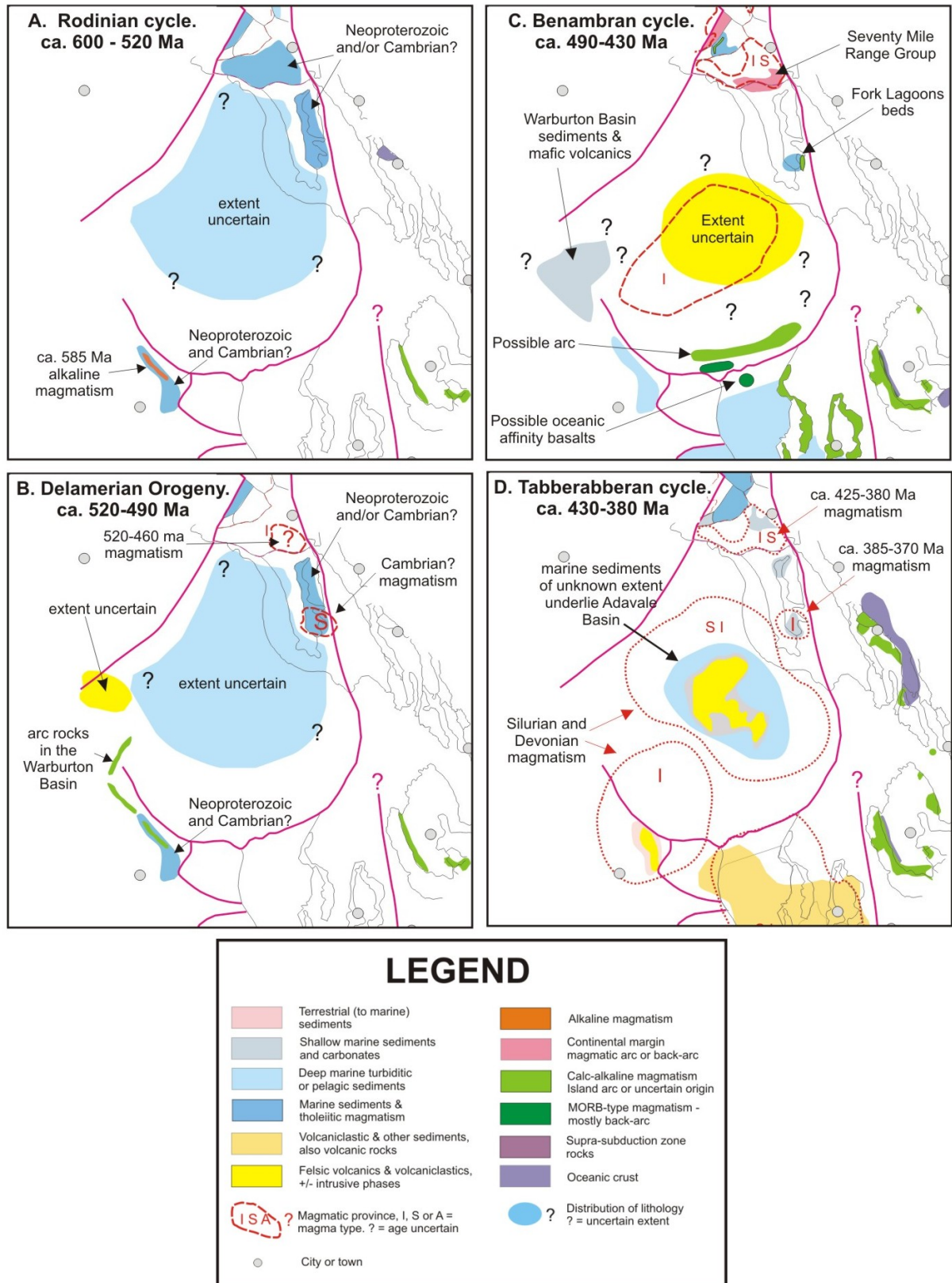


Figure 21. Generalised distribution of rocks in the Thomson Orogen, Koonenberry region and cover basins, by tectonic cycle. A = Rodinian, B = Delamerian, C = Benambran, D = Tabberabberan.

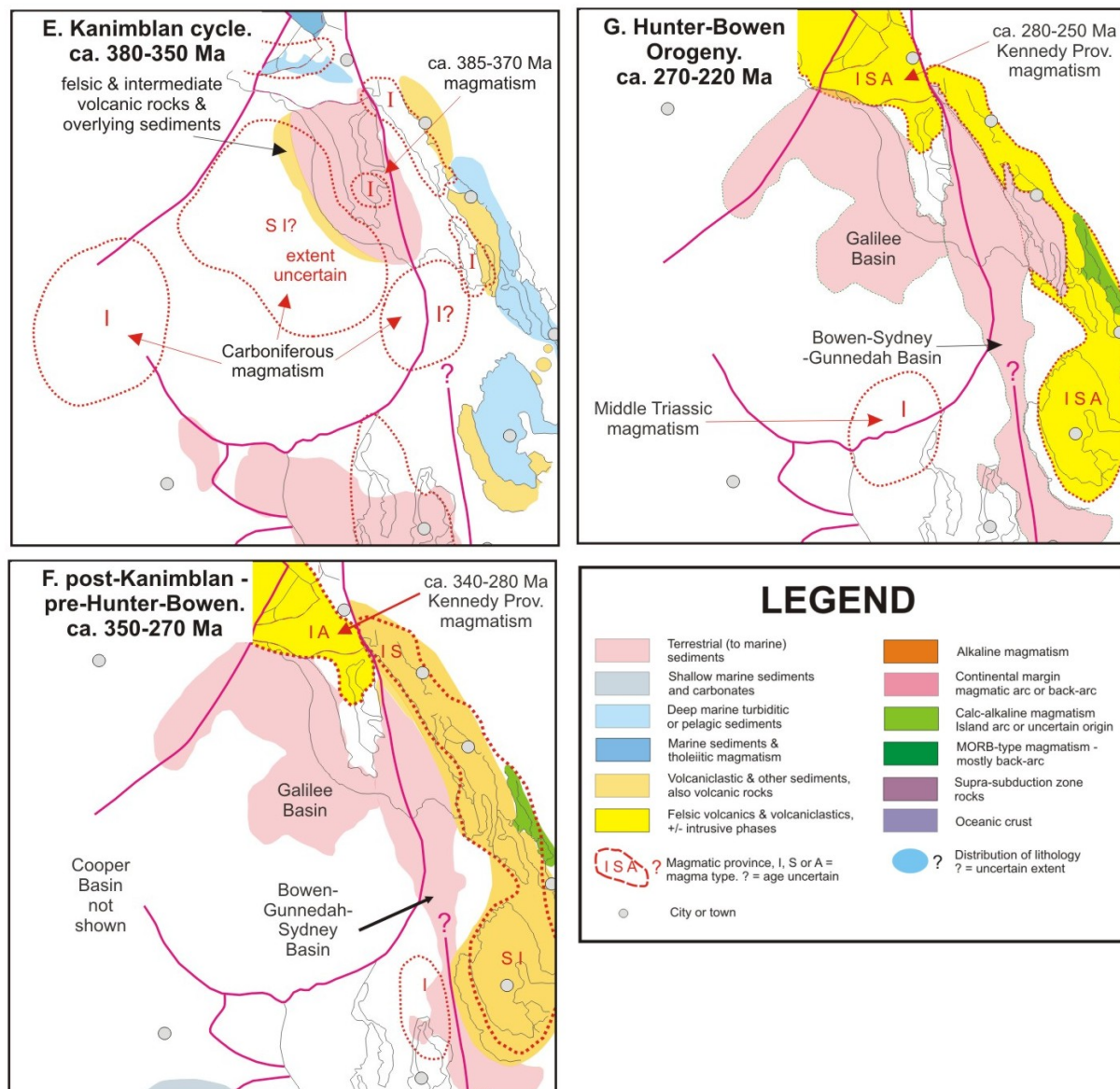


Figure 21 continued. Generalised distribution of rocks in the Thomson Orogen, Koonenberry region and cover basins, by tectonic cycle. E = Kanimblan, F = post-Kanimblan – pre-Hunter-Bowen, G = Hunter-Bowen.

- Tholeiitic and/or alkaline magmatism, consistent with rifting and/or passive margin volcanism is present within the Anakie Inlier (mafic schist), Cape River Metamorphics (amphibolite) and a high-Nb mafic volcanic suite in the Koonenberry Belt (Withnall et al., 1995, 1997; Crawford et al., 1997; Hutton et al., 1997; Fergusson et al., 2007; [Figs 20, 21](#)).

Central Thomson Orogen

- Although isotopic dating is limited, metasedimentary rocks within the central Thomson Orogen have been tentatively correlated with late Neoproterozoic to Middle Cambrian rocks in the Anakie Inlier and the Charters Towers region (Draper, 2006; [Figs 20, 21](#)). These rocks are overlain by subhorizontal Early Ordovician volcanic rocks and so must have been deformed and metamorphosed prior to this (Draper, 2006). Murray (1994) documents (non-SHRIMP) ages from basement cores elsewhere in the Thomson Orogen that suggest that Neoproterozoic to Middle Cambrian rocks may be more widespread in the Orogen.

Koonenberry Belt

- Sediments and volcanics of the Grey Range Group (continental shelf to deep marine) and equivalent Farnell Group, formed during intracontinental rifting associated with Rodinian break-up (Crawford et al., 1997; Gilmore et al., 2007; [Figs 20, 21](#)).
- The Mount Arrowsmith Volcanics (Grey Range Group) (alkali basalt, trachybasalt, trachyte, submarine and subaerial lavas, pyroclastics and related intrusives) are dated at ca. 585 Ma (Crawford et al., 1997; Black, 2007; [Figs 20, 21](#)). Crawford et al. (1997) interpreted these rocks as having been generated in a continental rift environment. Ultramafic and mafic rocks (peridotites and gabbros) were also formed at this time although their exact age and relationship to the Mount Arrowsmith Volcanics is unknown (Crawford et al., 1997).
- Sediments of the Gnalta Group (shallow marine/shelf siltstone and sandstone), Teltawongee Group (continental slope turbidites), and tholeiitic submarine volcanics of the Ponto Group were deposited along the passive continental margin during Cambrian extension (~542 to ~505 Ma) (Gilmore et al., 2007; [Figs 20, 21](#)). Tuffs in the Ponto Group have been dated at ca. 512 Ma and 509 Ma (Black, 2005). In the upper Gnalta Group, an ignimbrite has been dated at ca. 511 Ma (Black, 2007). Calcalkaline volcanism (andesite, basalt, rhyolite and dacite) within the Mount Wright Volcanics were also deposited at this time (Crawford et al., 1997). Gilmore et al. (2007) suggest either a backarc or forearc environment for these rocks. Crawford et al. (1997) suggested formation was in an immature continental rift, evolving to later tholeiitic magmatism with more extension.

1.4.2. Early to Middle Cambrian

Delamerian Orogeny: ca. 520 to ca. 490 Ma

Geological and tectonic summary

The Delamerian Orogeny deformed and metamorphosed rocks of the Anakie Inlier, Charters Towers region, Koonenberry Belt and central Thomson Orogen, and generally resulted in downwarping of the Warburton Basin (Murray and Kirkegaard, 1978). It is best recorded in basement rocks of the Anakie and Koonenberry regions. Based on comparative evidence from the Anakie Inlier, Draper (2006) suggested that deformation of subsurface metasedimentary rocks in the eastern Thomson Orogen was likely to have occurred during the Delamerian Orogeny. The contractional event was predominantly east-west in the Thomson Orogen, and northwest-southeast in the Koonenberry Belt (Gilmore et al., 2007). Uplift and erosion of the Warburton Basin was coincident with the Delamerian Orogeny (locally recognised as the Mootwingee Movement; Gravestock and Gatehouse, 1995), producing a brecciated interval and

depositional hiatus (Boucher, 2001; Harvey and Hibburt, 1999). This event also deformed the Adelaide and Kanmantoo Orogens, and broadly similar ages have been recognised in the Koonenberry, Charters Towers and South Australia, although deformation appears to have been shorter-lived in northern Australia (Foden et al., 2006; Black, 2007; Fergusson et al., 2007a, b).

Delamerian deformation was accompanied by syn- to post-orogenic calcalkaline magmatism in the Anakie Inlier and Koonenberry Belt (Withnall et al., 1995; Crawford et al., 1997; Gilmore et al., 2007; [Figs 20, 21](#)), and in the Warburton Basin (Gatehouse, 2006). As suggested by Gatehouse (1986), the calcalkaline volcanism may relate to the presence of an arc at this time. Although the Bourke-Louth regions are adjacent to the Koonenberry Belt, rocks of this age do not appear to have been recorded. The possible presence of an arc in the Warburton-Koonenberry region at this time, well west of the Anakie Inlier, is problematical. Withnall et al. (1995) suggested this may have either reflected a very wide Delamerian Orogen or subsequent (post-Delamerian) extension and rifting of the Anakie Inlier eastwards. The latter is not consistent, however, with the discovery of subhorizontal Ordovician volcanic rocks overlying deformed and metamorphosed metasedimentary rocks in the central Thomson Orogen (Draper, 2006; see below).

Geological history and Time-Space plot explanation

Anakie Inlier (Anakie region)

- The Anakie Metamorphic Group was complexly deformed and metamorphosed, to greenschist to amphibolite facies, in the Middle Cambrian, ca. 500 Ma (Withnall et al., 1995; 1996; Green et al., 2008; Fergusson et al., 2007c). Withnall et al. (1996) suggested that this deformation represented the northern continuation of the Delamerian Orogen.
- Syn- or post-orogenic S-type granites (Mooramin and Gem Park Granites), possibly of Cambrian (or Ordovician) age, occur within the Anakie Inlier (Crouch et al., 1995, Withnall et al., 1995; [Figs 20, 21](#)).

Koonenberry Belt

- Contractional, south-southeast – north-northwest, deformation and associated low-grade regional metamorphism was recognised by Gilmore et al. (2007) in the Koonenberry region. Deformation resulted in tight, west-vergent folding and thrusting with a component of sinistral strike-slip movement (Gilmore et al., 2007). The Williams Peak Granite provides the best evidence for the onset of the Delamerian Orogeny. It has been dated at ca. 516 Ma and indicates intrusion was pre- to synkinematic (Black, 2007). The end of Delamerian deformation is marked by a felsic intrusive cross-cutting the Delamerian foliation, and has been dated at ca. 497 Ma (Black, 2005).
- Volcanism in the Mount Wright Volcanics and Gnalta Group is linked to the Delamerian Orogeny ([Figs 20, 21](#)). Calcalkaline volcanism and tholeiitic basalts (MORB affinities) within these rocks have been used to suggest either forearc or backarc environments (Gilmore et al., 2007), or continental rifting (Crawford et al., 1997).

Thomson Orogen

- Middle Cambrian volcanism is recorded at ca. 510 Ma (Draper, 2006) in basement rhyolitic ignimbrite beneath the Poolowanna Trough and Eromanga Basin (DIO Adria Downs-1 well, north of the Queensland-South Australia border; [Figs 20, 21](#)), suggesting a link with deformation. The relationship of these volcanics to the Mooracoochie Volcanics (trachyte and dacites) in the Warburton Basin is unresolved due to overlying sedimentary basin cover.

Warburton Basin

- Fracturing of the existing Proterozoic craton led to arc-related volcanism and eruption of the Mooracoochie Volcanics in the Early Cambrian along what has been called the Gidgealpa Volcanic Arc (Gatehouse, 1986; Boucher, 2001; [Figs 20, 21](#)). Porphyritic trachyte and dacitic lavas from the Mooracoochie Volcanics have been dated at ca. 517 Ma (Malgoona-1 well; U-Pb zircon; Boucher, 2001). This date correlates with the U-Pb age of ca. 510 Ma for an unnamed rhyolitic ignimbrite beneath the Cooper Basin (Adria Downs-1 well; Draper, 2006), and suggests these two units are equivalents (Draper, 2006). The Mooracoochie Volcanics are also thought to be correlatives of the Cambrian volcanics in the Koonenberry Belt (Gravestock and Gatehouse, 1995), lending support to the arc model of Gilmore et al. (2007).
- Delamerian uplift and erosion (i.e., Mootwingee Movement of Gravestock and Gatehouse, 1995) produced a depositional hiatus adjacent to the Australian craton prior to the onset of marine sedimentation (e.g., Dullingari Group).

1.4.3. Middle Cambrian to Ordovician-earliest Silurian

Post-Delamerian to Benambran Orogeny: ca. 490 to ca. 430 Ma

Geological and tectonic summary

Middle Cambrian, Ordovician and Early Silurian rocks in the Thomson Orogen are dominated by shelf and deltaic sedimentary deposits (Koonenberry, Fork Lagoons), backarc volcanism (Fork Lagoons, Louth), and magmatism (Thomson, Fork Lagoons; [Figs 20, 21](#)). The magmatism of this age forms part of the widespread Macrossan Province in northern Queensland (e.g., Hutton et al., 1997; see north Queensland section; [Figs 20, 21](#)). Elsewhere the Late Cambrian to Ordovician marks the onset of more widespread marine basin sedimentation following the Delamerian Orogeny and intracratonic rifting (i.e. Warburton Basin; Gatehouse and Cooper, 1986). Progressive marine incursion southwards caused an expansion of basins across central Gondwana as a shallow epicontinental sea (Larapintine Sea), believed to link the Warburton, Amadeus and Canning Basins (Gravestock and Gatehouse, 1995; Maidment et al., 2007). This seaway may have also connected basinal environments in the Koonenberry region, although this is yet to be proven. Sediments were deposited in shelf and trough environments in the Warburton Basin, before being interrupted by the Benambran (~Alice Springs (1)) Orogeny which caused uplift and associated regression of the Larapintine Sea (Gravestock and Gatehouse, 1995).

The Early Silurian contractional Benambran Orogeny is represented in the Charters Towers region, Warburton Basin and Koonenberry Belt. Based on metamorphic ages (e.g., Draper, 2006), it is presumed that the Benambran event deformed rocks of the Thomson basement, but the areal extent of Benambran deformation in the Thomson is yet to be resolved.

In the Thomson Orogen magmatism took place in the Early Ordovician, Middle Ordovician and Middle Silurian (Murray, 1994; Draper, 2006; [Figs 20, 21](#)). Mafic to felsic magmatism of the Macrossan Province is best developed in the Charters Towers region, which is dominated by I-type and mantle magmatism with ages from ca. 490 Ma to ca. 455 Ma (e.g., Hutton et al., 1997). Early to Middle Ordovician volcanic- or volcanoclastic-dominated successions in the Charters Towers region have a calcalkaline signature, and are interpreted as having formed in a backarc environment (e.g., Seventy Mile Range Group; Henderson, 1986; Stolz, 1994).

Geological history and Time-Space plot explanation

Anakie region

- Late Ordovician marine metasedimentary rocks including carbonates and associated mafic to intermediate volcanic rocks of the Fork Lagoons beds (Withnall et al., 1995; [Figs 20, 21](#)). Intrusion

of gabbro was comagmatic with basalt and andesite extrusion; the geochemistry of these rocks suggests tholeiites of either island arc or backarc origin (Withnall et al., 1995). The Fork Lagoons beds also contain structurally emplaced serpentinite (Withnall et al., 1995). The metasedimentary rocks of the Fork Lagoons beds were apparently sourced from both cratonic and volcanic provenances (e.g., Fergusson et al., 2007c), which led Withnall et al. (1995) to suggest an arc setting not too far from a continent. Withnall et al. (1995) also suggested that the relationship between the Fork Lagoons beds and the older Anakie Metamorphic Group was similar to that observed between early Palaeozoic and Neoproterozoic rocks in the Georgetown region.

- Poorly constrained northwest-southeast contractional deformation in the Fork Lagoons beds may relate to either Benambran or Tabberabberan deformation (Withnall et al., 1995).

Koonenberry Belt

- Cessation of Delamerian contraction, in the Late Cambrian to Early Ordovician, resulted in uplift and erosion to produce local extensional basins, with deposition of shelf, to deltaic and deep-water sediments of the Kayrunnera, Mutawintji and Warratta Groups (Gilmore et al., 2007; [Figs 20, 21](#)).
- Contractional deformation associated with the Benambran Orogeny occurred in the Late Ordovician to Early Silurian (Gilmore et al., 2007). According to Gilmore et al. (2007) deformation was most intense east of the Koonenberry Fault. These authors also record a vergence change across this fault – from east-vergent in the east, to west-vergent west of the fault.

Bourke-Louth regions

- Early Ordovician to Early Silurian backarc volcanism and associated sedimentation has been recognised in the Louth (Thomson Orogen) and nearby Mount Dijou (Lachlan Orogen) regions (northwest NSW) (Watkins, 2007; [Figs 20, 21](#)). Preliminary dating of these units, summarised by Watkins (2007), give an Early Ordovician age (ca. 484 Ma) for the Louth volcanoclastics (also supported by fossil evidence), and an Early Silurian age (441 Ma) for mafic-intermediate volcanics. The rocks of both regions are characterised by both calcalkaline to shoshonitic, and alkaline (with OIB-like patterns), compositions (Watkins, 2007; Burton et al., 2008; [Fig. 21](#)). The calcalkaline compositions appear to be similar chemically to the Ordovician Lachlan Macquarie Arc (Burton et al., 2008), and Watkins (2007) suggested the possible presence of a contemporaneous oceanic arc in the Thomson Orogen. Watkins (2007) appears to suggest a south-dipping slab, effectively putting the alkaline rocks, which occur south of the calc-alkaline rocks ([Fig. 21](#)) in a backarc position.

Thomson Orogen

- Felsic volcanism and granite intrusion occurred in the eastern and central Thomson Orogen (Murray, 1994), and has been dated as Early Ordovician to Middle Silurian (Draper, 2006; [Figs 20, 21](#)). Porphyritic rhyolite, rhyolitic tuff and crystal tuff form Thomson basement beneath the Eromanga (GSQ Maneroo-1), Drummond (BEA Coreena-1) and Adavale (Carlow-1) basins and have been dated as Early Ordovician (473, 478, and 484 Ma; McKillop et al., 2005; Draper, 2006).
- Felsic magmatism occurred in the western Thomson Orogen beneath the Eromanga and Cooper basins (AMX Toobrac-1; DIO Ella-1; Murray, 1994). Recent age dating (reported in Draper, 2006) indicates Middle Ordovician (470 Ma) and Middle Silurian (428 Ma) ages. Similar Early to Middle Ordovician ages of the volcanics imply that the volcanic and plutonic rocks are closely related and may be comagmatic during extension and crustal thinning during the Ordovician (Draper, 2006).
- Sedimentary rocks of the Adavale Basin are underlain by Lower Ordovician basement volcanics which have been dated at ca. 484 and 489 Ma (McKillop et al., 2005; [Figs 20, 21](#)).

Warburton Basin

- Following the Delamerian Orogeny and basin rifting and volcanism, shallow shelf, slope and basin sediments of the Kalladeina Formation (carbonates) and deep water sediments of the Dullingari Group (shale, siltstone and chert) were deposited along the margins of the Australian craton (Gatehouse and Cooper, 1986; Boucher, 2001; [Figs 20, 21](#)).

- Rift-related, continental/intraplate mafic volcanics (Jena Basalt) and related agglomerate were extruded in the lower part of the Dullingari Group in the Middle Cambrian (Meixner et al., 1999, 2000; Boucher, 2001).
- Red beds of the Innamincka Formation were deposited in a shallow marine epicontinental sea and although they have been drilled extensively, their age is unrestrained (Gatehouse, 1986; Figs 20, 21).
- Sediments underwent Early Silurian Benambran/Alice Springs Orogeny (1) folding and erosion (Gatehouse, 1986). Cambrian and Late Ordovician rocks were probably metamorphosed in the Silurian (Draper, 2006), although further work is necessary to determine the precise age.

1.4.4. Middle Silurian to Middle to early Late Devonian

Post-Benambran to Tabberabberan Orogeny: ca. 430 to ca. 380 Ma

Geological and tectonic summary

During this cycle, both terrestrial and marine sedimentation and associated extrusive and intrusive magmatism occurred, within two episodes, largely in response to extension following the Benambran and Bindian orogenies (e.g., Olgers, 1972; Neef and Bottrill, 1991; Murray, 1994, 1997a; Withnall et al., 1995; McKillop et al., 2005; Gilmore et al., 2007; Figs 20, 21). Post-Benambran, Late Silurian to Early Devonian deposition is recorded in the Koonenberry region (Neef and Bottrill, 1991), and in the Thomson Orogen (Murray, 1994; 1997a; Withnall et al., 1995). These consist of marine and terrestrial sediments, of cratonic and/or volcanic provenance. They are associated with mafic and felsic volcanic rocks in the Koonenberry region (Neef and Bottrill, 1991) and in the Anakie Inlier (Withnall et al., 1995), and felsic intrusives in the Thomson Orogen basement (e.g., Murray, 1994) and the Koonenberry region (Gilmore et al., 2007). Both Murray (1994) and Thalhammer et al. (1998) have suggested continental settings. Widespread felsic intrusive magmatism of this age (ca. 425–405 Ma), belonging to the Pama Magmatic Province (Bain and Draper, 1997) occurs within the Charters Towers region (Hutton et al., 1997). Local folding and metamorphism of these successions, probably with associated felsic magmatism, took place during the Bindian Orogeny (e.g., Thalhammer et al., 1998). The orogeny may have been diachronous. Deformation in the Koonenberry region is recorded as Late Silurian to Early Devonian (Gilmore et al., 2007), while it appears to be late Early Devonian in the Thomson Orogen, where it is constrained by ca. 408 and 402 Ma volcanic rocks in the post-Bindian Adavale Basin (McKillop et al., 2005).

Renewed extension, following the Bindian Orogeny, produced the Early to Late Devonian, terrestrial to shallow marine, Adavale Basin in the central Thomson Orogen (McKillop et al., 2005), terrestrial to shallow marine sedimentation in the Burdekin Basin (Charters Towers region; Hutton et al., 1997), and Early to Middle Devonian quartz-rich sedimentation in the Koonenberry region (Neef, 2004). The Adavale Basin is thought to have formed in a continental setting, either as an intracontinental volcanic rift or an extensional basin (Murray, 1994; McKillop et al., 2005). Felsic intrusive magmatism accompanied extension in the Thomson Orogen (e.g., Murray, 1994; Evans et al., 1990; Hutton et al., 1997; Figs 20, 21). McKillop et al. (2005) suggest that extension may have been the result of more regional events such as subduction further to the east in the New England Orogen. The Middle Devonian Tabberabberan Orogeny (which has been called the Alice Springs Orogeny (2) in the Thomson Orogen) is best recorded in the Koonenberry region where it resulted in east-northeast – west-southwest contractional deformation at ca. 395 Ma (Mills and David, 2004; Neef, 2004). Deformation of this age in the Thomson Orogen is either poorly developed or difficult to distinguish from other events (Withnall et al., 1995; Hutton et al., 1997). In the Adavale Basin, the orogeny appears to have resulted in an unconformity between terrestrial and overlying shallow marine sedimentary rocks, and a possible change to restricted basin conditions in the late Middle Devonian (McKillop et al., 2005).

Geological history and Time-Space plot explanation

Anakie region

- Deposition of carbonates and clastic sediments, with associated mafic and felsic volcanic rocks, in the Early to Middle Devonian Douglas Creek Limestone and Glendarriwell beds (Olgers, 1972; Withnall et al., 1995; [Figs 20, 21](#)). These were considered to be correlatives of the Ukalunda beds by Withnall et al. (1995).
- The Middle Devonian (ca. 385-370 Ma) Retreat Batholith (I-type) intrudes the Anakie Metamorphic Group (Webb and McDougall, 1968; Withnall et al., 1995), and post-dates regional metamorphism and probably the Tabberabberan Orogeny.

Drummond Basin (Anakie region)

- In the Early Devonian, isolated marine deposition of siliciclastics (Ukalunda beds) were deposited on top of metasedimentary rocks in the Anakie Inlier, and granites and volcanics of the central Thomson Orogen basement (Olgers, 1972; Grimes et al., 1986; Murray, 1994; [Figs 20, 21](#)). The Ukalunda beds form basement to the Drummond Basin (Grimes et al., 1986; Hutton et al., 1998; Draper et al., 2004).

Adavale Basin

- Felsic volcanic and volcanoclastic rocks (acid crystal tuffs, ignimbrite, minor mafics) of the Gumbardo Formation formed during initial rifting and half graben formation in the late Early Devonian (McKillop et al., 2005). The volcanics have been dated as late Early Devonian (402 Ma, 408 Ma; Gumbardo-1 and Carlow-1 wells; McKillop et al., 2005), and were deposited in fluvial or fluvial-lacustrine conditions (McKillop et al., 2005).
- Terrestrial, fluvial-lacustrine and marginal marine sedimentation (Eastwood Formation) continued until the Late Devonian. A marine transgression produced a phase of shallow marine carbonate deposition (Log Creek Formation, Bury Limestone) from the Early to Middle Devonian Alice Springs Orogeny (2) (McKillop et al., 2005).
- The Alice Springs Orogeny (2) (~Tabberabberan Orogeny) is thought to have produced an unconformity between terrestrial and overlying shallow marine sedimentary rocks, and may have triggered the onset of restricted basin conditions in the late Middle Devonian (mid Givetian), which produced a sabkha environment with associated halite deposition (McKillop et al., 2005).

Thomson Orogen

- Widespread Late Silurian to Early Devonian siliciclastic (marine) sedimentation (quartz turbidites) were deposited prior to the overlying Adavale Basin (Murray, 1994). The sedimentary character of these rocks is similar to the Timburry Hills Formation (Murray, 1994, p.79). Folding and metamorphism of these rocks may have occurred shortly after cessation of deposition (Murray, 1994), probably related to the Bindian Orogeny(?).
- Middle Silurian and Early Devonian felsic intrusion into Thomson basement rocks ([Figs 20, 21](#)). These include an early phase of ?post-orogenic, granites (ca. 428 Ma), and a younger phase of granites (ca. 408-405 Ma; Draper, 2006). Murray (1994) suggested the latter granites may be related to extension of the Adavale Basin system.
- Late Middle Devonian (385 Ma) felsic volcanism (rhyolitic ignimbrite; AAE Towerhill-1; Draper, 2006) has been suggested to correlate with the Silver Hills Volcanics in the Drummond Basin (Murray, 1994). If correct, this would suggest the former are post-Tabberabberan, that is, belong to the Kanimblan Cycle.

Bowen Basin

- Devonian(?) deep marine (turbidite?) deposition of metasedimentary rocks of the Timburry Hills Formation (basement to Bowen Basin). The quartz sandstone unit is uniform in character and suggests a continental source (Murray, 1997a). A Devonian age is constrained by the presence of

plant material and emplacement of the Roma Granites close to the Devonian-Carboniferous boundary (Murray, 1997a).

Koonenberry Belt

- Deposition in the Early Devonian of the nonmarine sediments of the Mount Daubeny Basin derived from a mixed cratonic and volcanic provenance (Neef and Bottrill, 1991; [Figs 20, 21](#)). Both Neef and Bottrill (1991) and Gilmore et al. (2007) record andesitic volcanism (ca. 425 Ma; Black, 2007) and intrusives within the succession, and Neef and Bottrill (1991) suggested a proximal volcanic source for the volcanoclastic component of the sediments.
- The Late Silurian to Early Devonian Bindian Orogeny affected the Koonenberry region and largely resulted in dextral strike-slip movement (Gilmore et al., 2007). Apparently associated with the deformation are ca. 427 to 420 Ma monzodioritic and I-type granite intrusions (Black, 2006; Gilmore et al., 2007). These have intruded the Koonenberry region, possibly extending into the Thomson Orogen (Thalhammer, et al., 1998). The latter include the fractionated (I-type) Tibbooburra Granodiorite, which has been dated at 421 Ma (Gilmore et al., 2007; 410 Ma (Rb-Sr) by Thalhammer et al., 1998). Thalhammer et al. (1998) suggested that the granite was emplaced in an intracontinental setting, syntectonic with local Early Devonian deformation, probably Bindian Orogeny (Gilmore et al., 2007). High level intrusion of Late Silurian-Early Devonian rhyolites occurred ca. 418-414 Ma (Black, 2006; Gilmore et al., 2007).
- Post-Bindian extension resulted in deposition of the Early and Middle Devonian terrestrial to shallow marine, quartz-rich sediments of the Wana Karnu Group (Gilmore et al., 2007). This includes the Snake Cave Sandstone of the Darling Basin, described by Neef and co-workers, e.g., Neef (2004).
- East-northeast – west-southwest contractional deformation of the Tabberabberan Orogeny, ca. 395 Ma (Mills and David, 2004, Neef, 2004).

1.4.5. Late Devonian to Early Carboniferous

Post-Tabberabberan to Kanimblan Orogeny (Alice Springs 3): ca. 380 – ca. 350 Ma.

Geological and tectonic summary

During this cycle both terrestrial and marine sedimentation, often with accompanying volcanism, occurred (e.g., Koonenberry region, Anakie Inlier, Thomson Orogen, Drummond Basin; [Figs 20, 21](#)), largely in response to intracratonic extension following the Tabberabberan Orogeny, but also backarc extension behind a Late Devonian to Early Carboniferous arc in the New England Orogen (e.g., Neef and Bottrill, 1991; Murray, 1994; Withnall et al., 1995; Henderson et al., 1998; Draper et al., 2004; McKillop et al., 2005; Gilmore et al., 2007). This backarc extension resulted in rifting and initiation of the Drummond basin in the latest Devonian (Henderson et al., 1998; [Fig. 21](#)). The Drummond Basin consists of a thick succession of continental and lesser marine sediments and volcanics (Olgers, 1972; Hutton et al., 1998; Henderson et al., 1998). These have been subdivided into three major tectonostratigraphic cycles ([Fig. 20](#)), separated by unconformities (e.g., Olgers, 1972). The lowermost cycle – cycle 1 (latest Devonian to Early Carboniferous) – consists of syn-rift related volcanic rocks and associated marine to terrestrial volcanoclastic sediments (Olgers, 1972; Henderson et al., 1998). Early Carboniferous Cycle 2 rocks consist of a thick succession of terrestrial (and local marine) sediments which reflect an abrupt end to volcanism and a switch to a cratonic provenance (Olgers, 1972). Cycle 3 rocks (also Early Carboniferous) reflect a return to volcanism, which continued episodically, with terrestrial sediments (Olgers, 1972).

Felsic, and lesser amounts of intermediate and mafic, magmatism accompanied extension episodically throughout this cycle. This includes intrusive and related extrusive magmatism in the Anakie Inlier, (syn? to) post-Tabberabberan Orogeny, but prior to formation of the Drummond Basin. During the Late

Devonian to Early Carboniferous, the Thomson Orogen also experienced regionally extensive felsic magmatism (e.g., Murray, 1994; [Figs 20, 21](#)). Early Carboniferous granites were emplaced into Thomson Orogen basement and the Warburton Basin (e.g., Murray, 1994). During initial Late Devonian-Early Carboniferous backarc extension, silicic magmatism at this time was spread over a broad region in the Drummond Basin and may be related to episodes of silicic magmatism in the New England Orogen (e.g., Bryan et al., 2004). This magmatism largely predates the widespread Kennedy Province magmatism in Charters Towers and further north, although volcanism in cycle 3 of the Drummond Basin appears to correlate with volcanism in the upper parts of the Burdekin, Bundock and Clarke River basins in the Charters Towers and Broken River regions of north Queensland (e.g., Henderson et al., 1998; [Fig. 14](#)).

The Early to ?Middle Carboniferous Kanimblan Orogeny, or Alice Springs Orogeny (3) as it is known in the Thomson Orogen, produced a major episode of faulting and deformation in the Koonenberry Belt, slight contraction in the Drummond Basin and regional-scale folding and subsequent erosion in the Adavale Basin (e.g., Olgers, 1972; Neef, 2004; Gilmore et al., 2007). This deformation event is suspected to have driven regional-scale, southward thrusting of the Thomson over the Lachlan Orogen (Korsch et al., 1997). Deformation in the Drummond Basin is recorded by an unconformity between the Drummond and Galilee Basin (Scott et al., in prep.).

Geological history and Time-Space plot explanation

Anakie Inlier (Anakie region)

- The Middle Devonian (ca. 385-370 Ma) Retreat Batholith (I-type) intrudes the Anakie Metamorphic Group (Webb and McDougall, 1968; Olgers, 1972; Withnall et al., 1995; [Figs 20, 21](#)), and post-dates regional metamorphism and probably the Tabberabberan Orogeny.
- Middle Devonian mafic to intermediate volcanics and associated volcanoclastics of the Theresa Creek Volcanics (Olgers, 1972; Withnall et al., 1995). The volcanics are locally intruded by granites of the Retreat Batholith and conformably overlie the Douglas Creek Volcanics (Olgers, 1972). Withnall et al. (1995) suggested that the volcanics were contemporaneous and probably genetically related to intrusive magmatism of the Retreat Batholith, most probably all generated in an extensional backarc environment, behind the New England Orogen.
- Early Late Devonian andesitic volcanism and minor terrestrial to marine sedimentation of the Greybank Volcanics (Withnall et al., 1995). Withnall et al. (1995) correlated these with the Dee Volcanics of the New England Orogen.
- Effects of the Kanimblan-Alice Springs Orogeny (3) in the Anakie Inlier is uncertain. Withnall et al. (1995) document minor folding and deformation in the Middle and Late Devonian volcanics and sediments that may be Kanimblan in age. Fenton and Jackson (1989) consider that Carboniferous deformation produced right-lateral offset and uplift of the Anakie Inlier as a series of blocks.

Drummond Basin (Anakie region)

- The Drummond Basin was initiated during this cycle – probably latest Late Devonian (Henderson et al., 1998). The basin consists of a thick succession of continental sediments and volcanics, with minor marine interbeds towards the base of the succession (Olgers, 1972; Hutton et al., 1998; Henderson et al., 1998; [Figs 20, 21](#)). The basin unconformably overlies the Early Devonian Ukalunda beds, the Retreat Batholith, and the older rocks of the Anakie Inlier (Olgers, 1972; Withnall et al., 1995). The Drummond Basin has been subdivided into three major tectonostratigraphic cycles, separated by unconformities (e.g., Olgers, 1972; Hutton et al., 1998; Scott et al., in prep; [Fig. 20](#)). The lowermost cycle – cycle 1 (latest Devonian to Early Carboniferous) – consists of syn-rift related volcanics rocks - andesitic, dacitic and dominant rhyolitic lava, ignimbrite and tuff - and associated marine to terrestrial volcanoclastic sediments (Olgers, 1972; Henderson et al., 1998). Henderson et al. (1998) indicated that volcanism and deposition initiated first in the north. Early Carboniferous Cycle 2 rocks consists of a thick sequence of terrestrial (and local marine) sediments – mostly quartz-rich sandstone, conglomerate and mudstone - which reflects an abrupt end to volcanism and a switch to a cratonic provenance

(Olgers, 1972). Cycle 3 rocks (also Early Carboniferous) reflect a return to volcanism, reflected by volcanoclastic sediment, tuff and conglomerate (Olgers, 1972). Volcanism continued episodically throughout this cycle, although the upper parts are dominated by terrestrial sediments. (Olgers, 1972). Most authors advocate an extensional backarc environment for the Drummond Basin (e.g., Henderson et al., 1998), with the arc situated within the New England Orogen.

- Deposition within the Drummond Basin was terminated by the Kanimblan Orogeny, which uplifted and folded the Drummond Basin (Olgers, 1972). This deformation is recorded by an unconformity between Drummond and Galilee Basin strata (Scott et al., in prep.).

Thomson Orogen

- Late Middle Devonian (385 Ma) felsic volcanism (rhyolitic ignimbrite; AAE Towerhill-1; Draper, 2006), which has been suggested to correlate with the Silver Hills Volcanics in the Drummond Basin (Murray, 1994).
- Widespread Late Devonian to Early Carboniferous (355-360 Ma) intrusion of S-type, and more localised I-type, Roma granites into the Timburry Hills Formation (Murray, 1994; [Figs 20, 21](#)). Mineralogy suggests the granites were intruded into old, stable continental crust (Murray, 1994).
- During north-south contraction of the Middle Carboniferous Alice Springs Orogeny (3), the Thomson Orogen was probably thrust over the Lachlan Orogen (Glen et al., 2007c).

Adavale Basin

- Termination of deposition and basin deformation occurred during Alice Springs Orogeny (3), with development of regional-scale folds followed by widespread erosion (McKillop et al., 2005).

Koonenberry Belt

- Post-Tabberabberan extension resulted in deposition of terrestrial quartz-rich sediments (e.g., the Ravensdale Formation), as part of the Darling Basin (e.g., Neef, 2004; [Figs 20, 21](#)).
- These rocks were deformed as part of the Kanimblan Orogeny, which produced regional faulting, fault reactivation as well as transpressive deformation (e.g., Neef, 2004; Gilmore et al., 2007).

1.4.6. Middle Carboniferous to Early Permian

Post-Kanimblan to orocline formation: ca. 350 to ca. 265 Ma

Geological and tectonic summary

From the Late Carboniferous to Early Permian, Eastern Australia was dominated by tectonic extension and rifting, which initiated extensive intracratonic basin formation (e.g., Cooper and Galilee basins; Bain and Draper, 1997; Korsch et al., 1998). Continental margin extension in the Early Permian to Middle Triassic led to formation of the Bowen and Gunnedah basins in a backarc setting (Korsch et al., in press; [Figs 20, 21](#)). Although extension was located on the continental margin, basins such as the Galilee and Cooper formed on the craton further west at this time (Draper and McKellar, 2002; [Fig. 21](#)). It has been suggested that deformation by Alice Springs Orogeny (3) caused convective downwelling and regional downwarp of the Drummond Basin, resulting in the formation of troughs and depressions in the Galilee Basin (Jackson et al., 1981; Middleton and Hunt, 1989). The Bowen Basin formed in an extensional environment east of the Drummond Basin during or almost immediately following Late Carboniferous batholith emplacement (Esterle and Sliwa, 2000). Age dating by Allen et al. (1998) suggests that the Bowen Basin formed in a backarc setting west of the Camboon Volcanic Arc in the New England Orogen. By the Early Permian, the Bowen Basin, together with the Gunnedah and Sydney Basins, formed the 'East Australian Rift System' (Korsch et al., 1998). Bimodal and andesitic volcanics at the base of the Bowen Basin suggests that Late Carboniferous to Early Permian crustal thinning was related to Bowen Basin rifting (Green et al., 1997b; Withnall et al., in prep.).

Although sedimentation had ceased in the Drummond Basin by the Late Carboniferous, relative tectonic stability in west and central Queensland during the late Carboniferous to Permian led to widespread terrestrial sedimentation in the intracratonic Cooper and Galilee basins (Draper, 2002a, b; 2004; [Figs 20, 21](#)). A connection between the Cooper and Galilee basins meant that the two basins experienced related sediment deposition (Scott et al., 1995). In the east, troughs created during early north-east to south-west extension of the Bowen Basin provided depocentres for initial terrestrial sedimentation and contemporaneous volcanoclastics (Green et al., 1997a; Korsch et al., 1998). Sedimentation in the Bowen Basin was contiguous with the Gunnedah Basin and both unconformably overlie the Drummond Basin (Green et al., 1997a; Scott et al., in prep).

Widespread Late Carboniferous to Early Permian igneous activity produced the Kennedy Province in the Charters Towers region ([Figs 14, 21](#)). Magmatism of this age is present further south in the Anakie, Drummond and Bowen regions, and also in the Warburton Basin (e.g., Gatehouse et al., 1995; Hutton et al., 1998; Denaro et al., 2004; Sliwa and Draper, 2005). This activity is the same age as Kennedy Province magmatism in northern Queensland (Bain and Draper, 1997). I-type plutons (e.g., Joe De-Little Granite, Billy-can Creek Granite) intruded the Anakie region and were comagmatic with silicic volcanics and caldera complexes of the Bulgonunna Volcanic Group (Oversby et al., 1994; Hutton et al., 1998).

Geological history and Time-Space plot explanation

Drummond Basin and northern Anakie Inlier (Anakie region)

- Eruption of the dominantly felsic Bulgonunna Volcanic Group in the Late Carboniferous to Early Permian(?) (ca. 305 Ma; Hutton et al., 1998). It is dominated by felsic extrusives including rhyolite, ignimbrite and minor related sediments (Olgers et al., 1972; Hutton et al., 1998). The group unconformably overlies the Mount Wyatt Formation (northern Drummond Basin; Olgers et al., 1972).
- Late Carboniferous comagmatic, and Permian multi-phase intrusions, smaller plutons and dykes occur on the margins of the Bulgonunna Volcanic Group, and have also intruded the Drummond Basin and Anakie Inlier (Olgers et al., 1972; Hutton et al., 1998; [Figs 20, 21](#)). Recorded ages range from ca. 308 Ma to ca. 287 Ma (Hutton et al., 1998; Scott et al., in prep).

Bowen Basin

- Late Carboniferous to Early Permian eruption of the basaltic, andesitic to rhyolitic Combarngo and Camboon Volcanics may be related to initial rifting and formation of the Bowen Basin (Green et al., 1997b). Bimodal compositions suggest they were erupted in a continental backarc or continental margin arc setting (Green et al., 1997b; Withnall et al., in prep).
- Eruption of the Late Carboniferous to middle Early Permian Lizzie Creek Volcanic Group, including the basaltic and andesitic Mount Benmore Volcanics and volcanoclastic rocks, rhyolite and dacite, prior to deposition of the Back Creek Group (e.g., Sliwa and Draper, 2005; Withnall et al., in prep.).
- Early Permian sediments of the Reids Dome beds were deposited in alluvial plain to lacustrine environments within the developing depocentres of the Taroom, Boomi and Denison troughs (Shaw, 2002; [Figs 20, 21](#)).

Gunnedah Basin

- Early to Late Permian marine deposition of the Maules Creek, Goonbri and Leard Formations and the Back Creek Group (Hamilton, 1993; Tadros, 1995; Shaw, 2002). The Lower Back Creek Group (Porcupine Formation) represents marine incursion and deposition in a transgressive fan delta complex and the Upper Back Creek Group (Watermark Formation) represents maximum marine transgression (Shaw, 2002).
- Early Permian eruption and deposition of thick basaltic and rhyolitic volcanic successions including the Boggabri Volcanics and Werrie Basalt, and equivalents in the Sydney Basin (e.g., Tadros, 1995).

Cooper Basin

- Early to Late Permian sediments of the Gidgealpa Group were deposited in glacial, fluvial and lacustrine environments with the Patchawarra Formation recording the waning stages of Early Permian glaciation (Gray and McKellar, 2002).
- Numerous, minor contractional events occurred during deposition of the Gidgealpa Group, and are thought to reflect episodes of the Hunter–Bowen Orogeny (e.g., Apak et al., 1997; Gray and McKellar, 2002).

Galilee Basin

- Late Carboniferous to Early Permian terrestrial (fluvial) sediment deposition of the glacial Joe Joe Group and Early Permian Aramac Coal measures (Scott et al., 1995; Figs 20, 21). Major coal deposition represents tectonic stability and expansion of basin depocentres in a fluvial (braided river) system.

Warburton Basin

- Middle Carboniferous (323 ± 5 Ma) and Early Permian (298 ± 4 Ma) felsic granite intrusion (Big Lake Suite) into the Warburton Basin (dated from Moomba-1 and McLeod-1; Gatehouse et al., 1995; Figs 20, 21). Emplacement could potentially be syntectonic with the Alice Springs Orogeny (3) of Central Australia.

1.4.7. Mid-Late Permian to Mid-Late Triassic

Post-Orocline formation to Hunter-Bowen Orogeny: ca. 265 to 230 Ma

Geological and tectonic summary

In general, tectonic stability continued throughout the Permian and into the Triassic and led to the widespread infilling of intracratonic basins. Fluvial and lacustrine systems were associated with extensive swamps in the Cooper and Galilee basins, which resulted in the continuous deposition of plant-rich material suitable for coal generation (Cowley, 2007; Figs 20, 21). In the Late Permian, coastal swamps formed in the subsiding Bowen Basin leading to an accumulation of extensive coal deposits (Shaw, 2002; Figs 20, 21). Sedimentation in the Bowen, Cooper and Galilee basins continued throughout the Permian and into the Triassic, until the Middle Triassic (Scott et al., 1995; Bain and Draper, 1997; Green et al., 1997b; Draper, 2002a, b).

At this time, sedimentary basins experienced the Hunter-Bowen Orogeny (e.g., Harrington and Korsch, 1985; Korsch et al., in press), which is proposed to have extended from the Middle Permian to the Middle or Late Triassic (ca. ~265 Ma to ~230 Ma; Korsch et al., in press). The New England Orogen was thrust westward during this event, which resulted in tectonic loading and subsidence in the Bowen and Gunnedah basins. The tectonic regime of these basins switched from initial extension, to contractional in the mid-Permian (Korsch et al., in press). In the mid-Permian, large volumes of volcanoclastic material were shed from the volcanic arc and deposited in the adjoining Bowen Basin during foreland loading (Green et al., 1997a). A well-developed mid Permian unconformity in the Bowen, Galilee and Cooper basins marks a change in both sedimentation style and tectonic regime (Bain and Draper, 1997). In the Bowen Basin, this change is characterised by east-west crustal extension and volcanic deposition (e.g., Lizzie Creek Volcanics) in the Early Permian, which underlies the unconformity. Above this unconformity, deposition is predominantly marine and highlights the start of contractional tectonics in the basin (Bain and Draper, 1997).

In the Late Triassic, a contractional event resulted in uplift and erosion, and the cessation of deposition in the Galilee, Cooper and Bowen basins (Apak et al., 1997; Korsch et al., 1998). This widespread event was

felt across eastern Australia and resulted in folding and uplift of parts of the Drummond Basin (Fenton and Jackson, 1989). Overall, the Permian-Triassic sedimentary successions underwent reactivation of existing structures, and both sedimentary and thermal subsidence (Draper and McKellar, 2002).

In the west, restricted magmatic activity occurred in the Cooper Basin and the Bourke-Louth region (Figs 20, 21). Localised basalts were generated in the Nappamerri Trough, suggesting a waning, but not completely diminished, thermal regime (Draper and McKellar, 2002). Draper (2002) proposed that there may be a link between the ongoing thermal activity implied by the Triassic or Early Jurassic basalts and basin-related subsidence; that is, subsidence of the Eromanga Basin was triggered by the same thermal regime that initiated subsidence in the Cooper Basin. Burton et al. (2007) suspect that the Midway Granite and coeval intrusives, east of Bourke, indicate a more spatially widespread Mid Triassic magmatic pulse than is currently recognised.

Geological history and Time-Space plot explanation

Drummond Basin (anakie region)

- Regional Middle Triassic (255-230 Ma) east-west contraction resulted in folding, thrusting, sinistral strike-slip movement and erosion (Olgers, 1972; Murray, 1990; Johnson and Henderson, 1991).

Cooper Basin

- Latest Permian to late Middle Triassic deposition of the Nappamerri Group, conformably above sediments of the Gidgealpa Group on an extensively vegetated, fluvial floodplain, with ephemeral lakes (Gray and McKellar, 2002).
- Late Triassic or Early Jurassic mafic volcanism in the southwestern part of the basin (Nappamerri Trough) produced olivine basalts with ages of 227 ± 3 Ma and 100 ± 9 Ma (Murray, 1994; Draper, 2002a, b).
- Major Late Triassic contraction produced widespread uplift and erosion following deposition of the Tinchoo Formation, and led to the cessation of sediment deposition (Apak et al., 1997; Korsch et al., 1998).

Galilee Basin

- Late Permian terrestrial sediment and coal deposition (Colinlea Sandstone, Bandanna Formation) in a major fluvial system with associated peat swamps (Scott et al., 1995; Figs 20, 21).
- Major east-west contraction in the Middle Permian resulted in uplift and erosion, and formation of a major unconformity, representing commencement of the Hunter-Bowen deformation (e.g., Draper, in Bain and Draper, 1997).

Bowen Basin

- During basin subsidence, widespread deposition of marine sediments of the Upper Back Creek Group occurred (Shaw, 2002).
- By the end of the Permian, peat-forming wetlands and associated fluvial systems led to the development of extensive coal measures (Shaw, 2002; Figs 20, 21).
- The onset of the Hunter–Bowen Orogeny is believed to correspond to a major unconformity surface within the Middle Permian Aldebaran Sandstone (Denison Trough area; e.g., Stephens et al., 1996; Korsch et al., 1998). This event resulted in the development of the Bowen Basin as a foreland basin.
- Uplift of the eastern arc, in the Late Permian, resulted in shedding of large quantities of volcanolithic alluvial sediments and terrestrial deposition of the Early Triassic Rewan Group and the Gunnedah Basin equivalent, the Digby Formation (Green et al., 1997a; Shaw, 2002). This event led to marine conditions contracting to the central western part of the Basin.

- Middle Triassic terrestrial (lacustrine) deposition of quartz sandstone in the Clematis Group reflects a change in sedimentary source from the eastern arc to the uplifted craton in the west (Fielding et al., 1990).
- Middle Triassic fluvial and lacustrine deposition of the uppermost Moolayember Formation. Volcanic sediments are present in the sequence and most likely originated from a volcanic arc source located in the east (Green et al., 1997a).
- Middle to late Triassic uplift and folding from regional compression, resulted in a termination of deposition, followed by erosion and peneplanation (Bain and Draper, 1997; Green et al., 1997).

Gunnedah Basin

- Late Triassic to ?Early Jurassic bimodal volcanism of the Garrawilla Volcanics (Glenrowan Intrusives, Bulga Complex) commenced prior to deposition of the overlying Surat Basin, around 218 Ma (Martin, 1993; Tadros, 1995). Volcanism produced subaerial mafic flows and pyroclastic deposits. This activity is spatially widespread and is suggested to have continued to the Early Cretaceous (e.g., Shaw, 2002).

Bourke-Louth regions

- In the Middle Triassic, magmatism produced highly fractionated I-type felsic intrusions (Midway Granite) and comagmatic quartz-feldspar porphyry dykes in the Bourke region (Figs 20, 21). The Midway Granite has been dated at 235 ± 1.4 Ma and is associated with tin and other base metal mineralisation (e.g., skarn-type Doradilla prospect; Burton et al., 2007).

SECTION 2. REGIONAL OVERVIEW OF THE TECTONIC DEVELOPMENT OF EASTERN AUSTRALIA IN THE PHANEROZOIC

by DC Champion and N Kositsin

Introduction

The geology and tectonic development of eastern Australia, particularly the Phanerozoic component – the Tasman Orogen (Scheibner and Veevers, 2000; Veevers, 2000, 2004; Cawood, 2005; Glen, 2005) - has been the focus of numerous studies, with a voluminous literature, including numerous orogen-based or more regional reviews (e.g., Murray, 1986; 1997; Murray et al., 1987; Coney, 1992; Seymour and Calver, 1995; Bain and Draper, 1997; Gray et al., 1997, 2003; Gray, 1997; Gray and Foster, 1997; 2004; Scheibner and Basden, 1998; Foster and Gray, 2000; VandenBerg et al., 2000; Veevers, 2000; Li and Powell, 2001; Crawford et al., 2003a; Glen, 2004; Cawood, 2005). The focus of this research has led to a plethora of tectonic models with perhaps the majority of differences focussed on the Lachlan Orogen (e.g., see Gray, 1997; Gray and Foster, 2004; [Figs 22, 24, 25](#)). Importantly, despite the differences, there is a general consensus that since the Late Neoproterozoic, eastern Australia has, broadly, been in three fundamental tectonic states:

- Rodinian-breakup (rifting) and ensuing passive margin – in its simplest form, basically formation of the Pacific Ocean; also corresponds broadly to development of the Australian component of the Gondwana margin (e.g., Li and Powell, 2001; Cawood, 2005).
- Alternating extensional and convergent orogenic cycles commencing in the Cambrian and continuing through to the Mesozoic, resulting in accretionary growth – as evidenced in the Lachlan, Thomson and New England orogens. Orogenic events include the Cambrian Delamerian through to the Permian-Triassic Hunter-Bowen orogenies. These cycles effectively ended with cratonisation, with the main arc system moving further offshore (arc rollback; e.g., Collins and Vernon, 1994; Jenkins et al., 2002; Collins and Richards, 2008; [Fig. 23](#)).
- Rifting and passive margin ($\pm$ hotspot activity), related to rifting of crustal fragments, and opening of ocean basins, especially related to Gondwana breakup (e.g., Veevers, 2004).

Although broadly simple, in detail the tectonic evolution of eastern Australia is clearly complex. Not only is there evidence for diachronous events (Gray et al., 2003), and possible arc switches (e.g., Murray, 2007), it is also clear that the current make-up of eastern Australian provinces (especially Palaeozoic to early Mesozoic) may represent an amalgamation of terranes that were not necessarily juxtaposed, that is, there are both allochthonous and autochthonous terranes. Controversy and uncertainty involves the original positions of potentially allochthonous blocks such as western Tasmania and the Selwyn Block (Cayley et al., 2002), the role of strike-slip movement (Willman et al., 2002), the presence of domains such as the Melbourne Zone which is missing evidence for major deformation events, as well as the nature of possible oceanic arc remnants (Macquarie Arc; Gamilaroi-Calliope, Jamieson?). This clearly also has important implications for mineralisation, for example, locating extensions to the mineralised Macquarie and Calliope island arcs. Another area of controversy concerns the actual positions of, and number of, arcs (if any), as best exemplified in the Ordovician and Silurian of the Lachlan Orogen (e.g., Wyborn, 1992; Gray, 1997; Soesoo et al., 1997; O'Halloran et al., 1998; Collins and Hobbs, 2001; Willman et al., 2002; Fergusson, 2003; Spaggiari et al., 2004; [Figs 24, 25](#)). More mundane but equally important controversies concern the actual location of arcs and discriminating between arc-forearc and backarc environments. Perhaps the best example of this is the (unresolved) debate regarding the interpretation of the sediments in the Hodgkinson Province (e.g., Henderson, 1980; Bultitude et al., 1993; 1997).

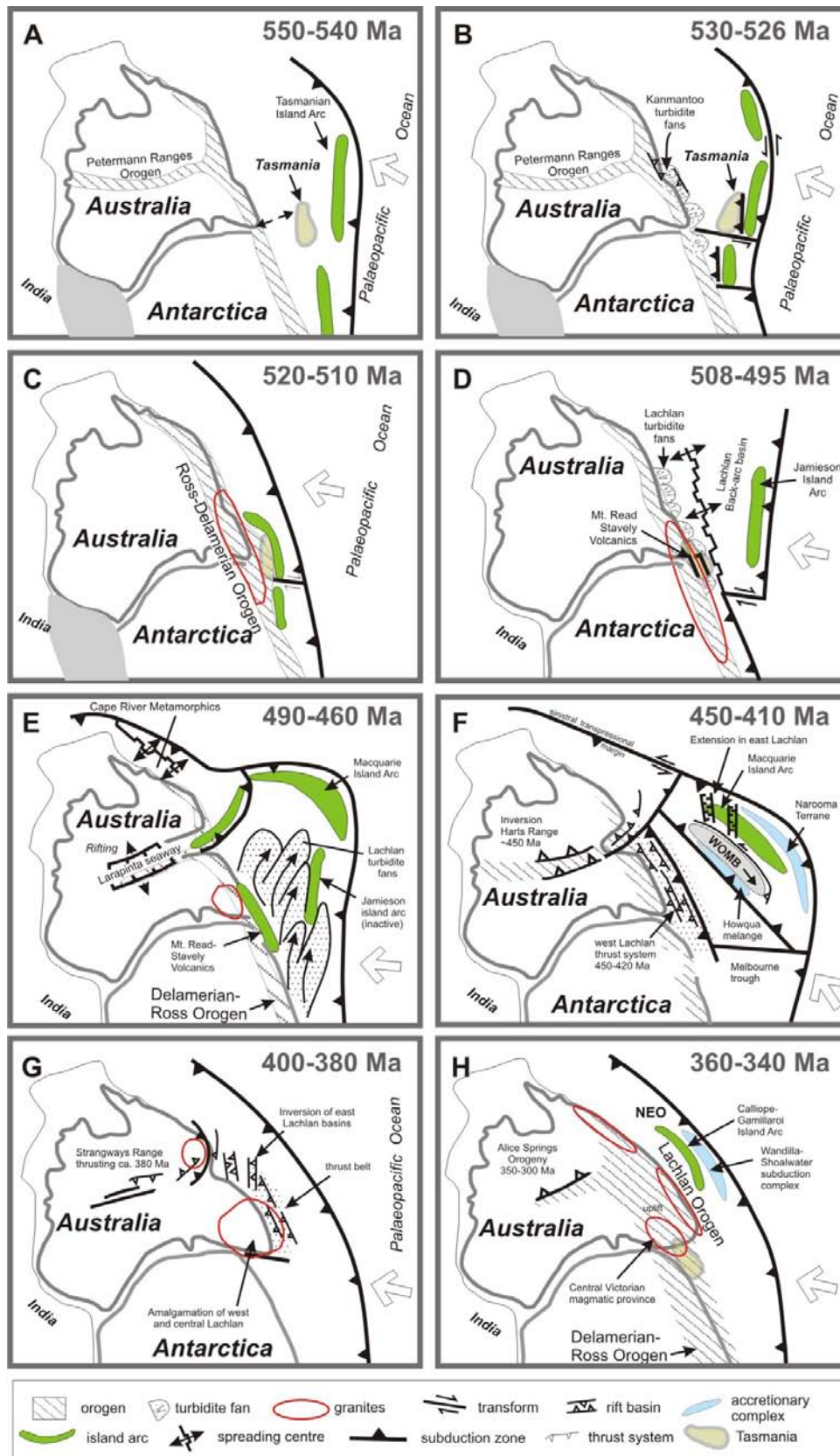


Figure 22. Tectonic evolution model of eastern Australia. Figure modified from Gray and Foster (2004).

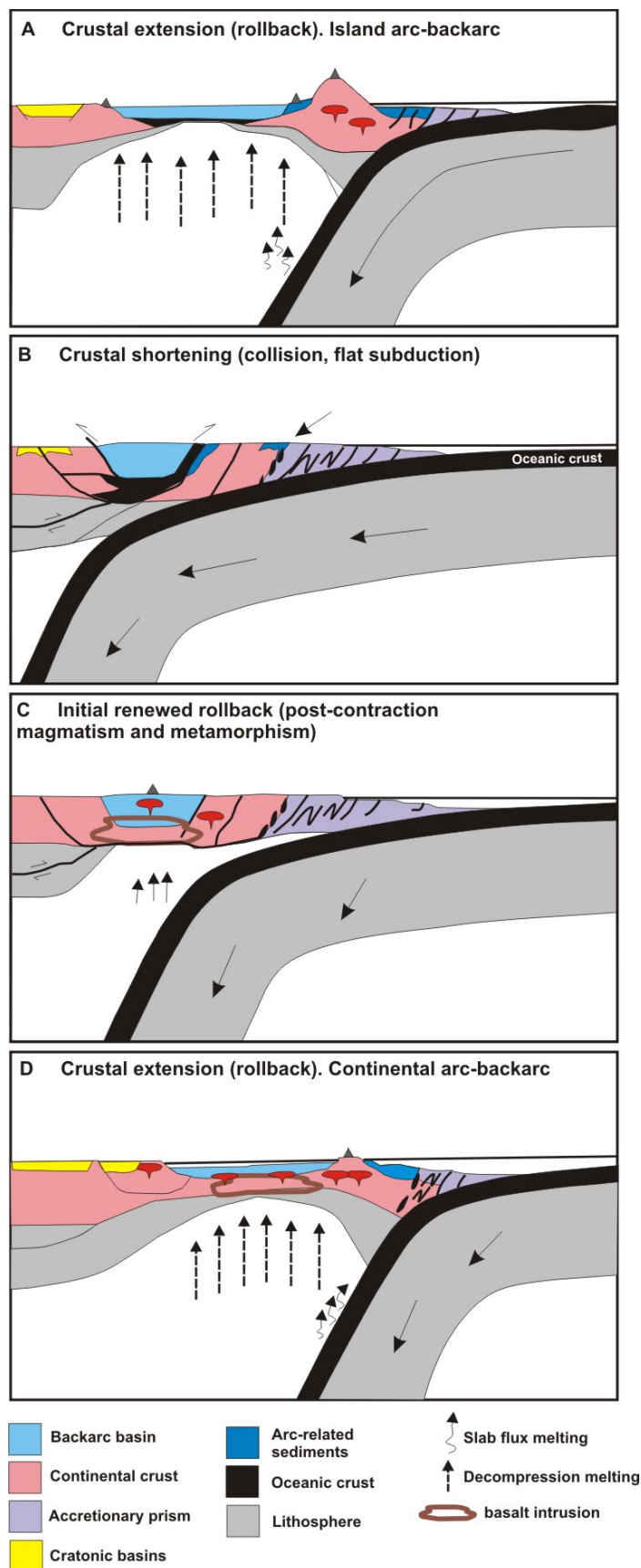


Figure 23. Interpreted tectonic environments for eastern Australia in the Palaeozoic, illustrating the cyclic alternation of extension and shortening for the interpreted orogenic cycles. Figure based on and modified from Collins and Richards (2008).

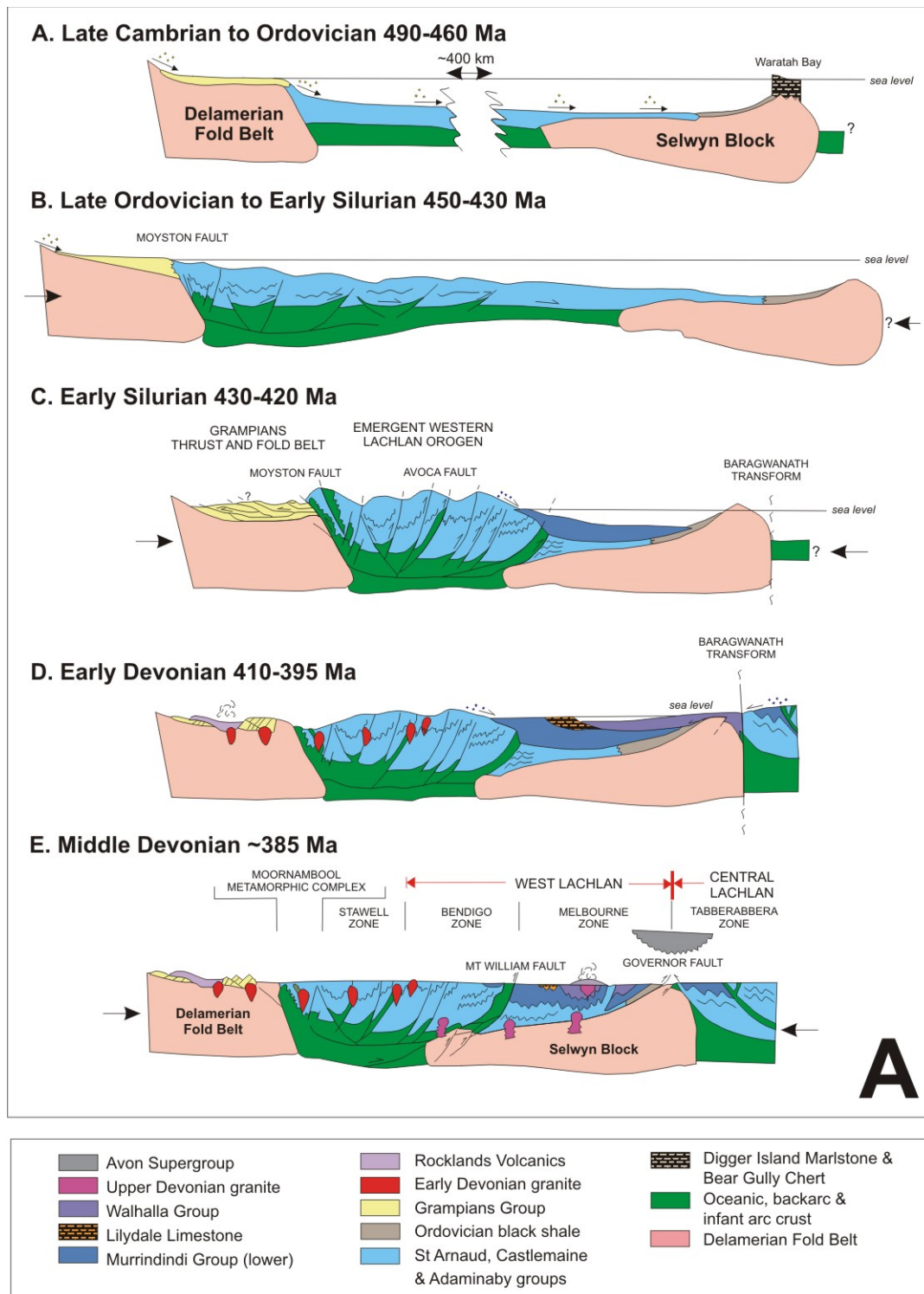


Figure 24. Schematic tectonic reconstructions for the Lachlan Orogen for the Ordovician to Devonian period. Tectonic model of VandenBerg et al. (2000) for the Western Lachlan Orogen in Victoria (Whitelaw Terrane). The reconstruction incorporates the Selwyn Block and the Baragwanath Transform, based on the models presented in VandenBerg et al. (2000), Cayley et al. (2002), and Willman et al. (2002). Figure based on (and modified from) from VandenBerg et al., 2000.

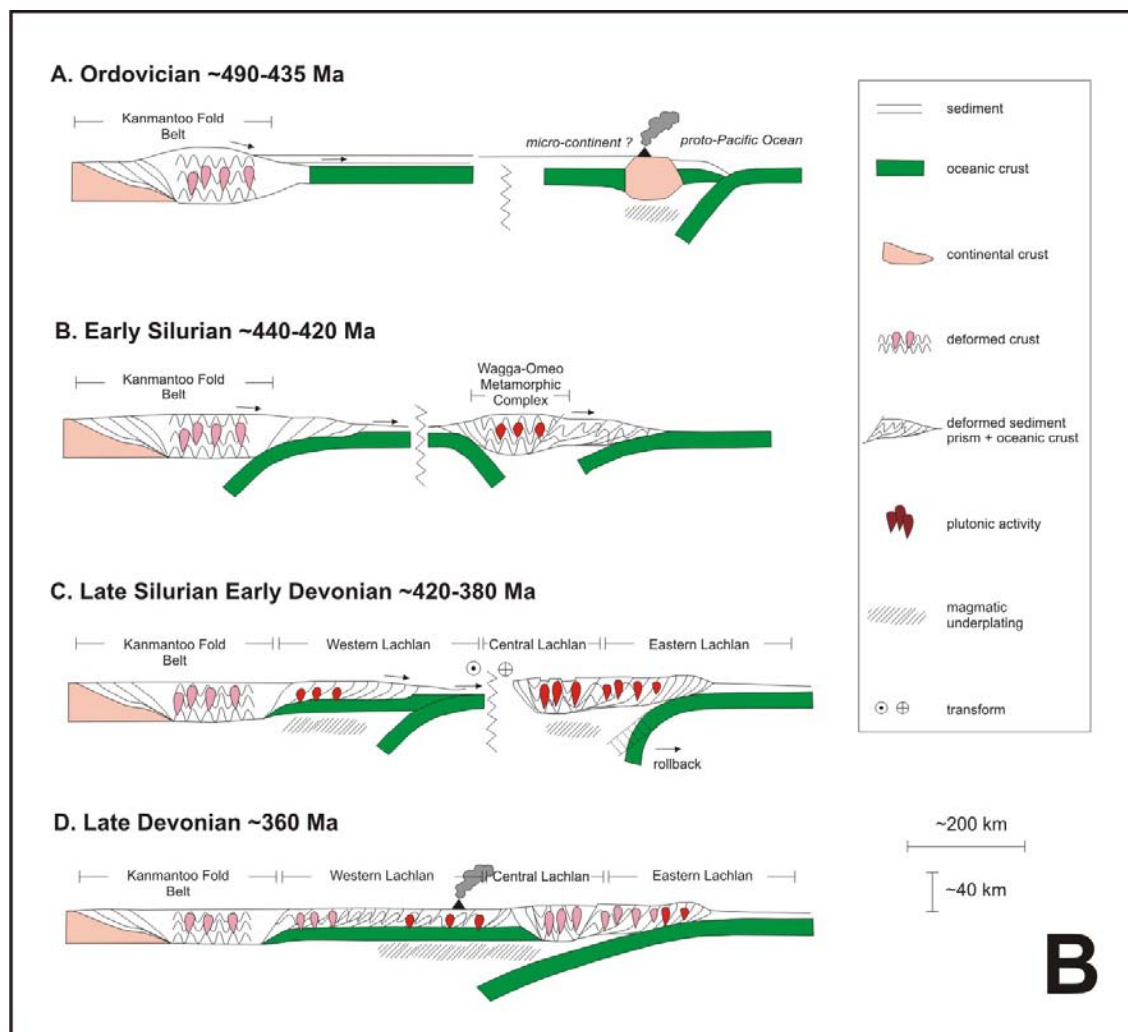


Figure 25. Schematic tectonic reconstructions for the Lachlan Orogen for the Ordovician to Devonian period. Tectonic model of Gray (1997) for the Western, Central and Eastern Lachlan Orogen, featuring the multiple subduction model presented by Gray and co-workers (e.g., Gray, 1997; Soesoo et al., 1997; Foster and Gray, 2004; Spaggiari et al., 2003). Figure based on (and modified from) from Gray (1997).

Most debate regarding tectonic reconstructions for eastern Australia, however, centre around the Lachlan Orogen, which although probably an accretionary margin, has, as summarised by Gray (1997), numerous features including its width, and variable deformation, that are not easy to explain by simple models (e.g., see Collins and Vernon, 1994). Further uncertainty concerns the Thomson Orogen, which is largely undercover, and as such poorly understood. Recent seismic results across the southern margin of this orogen suggest that this boundary is not simple: the MOHO thickens to the north and the boundary may represent collision between the Lachlan and Thomson orogens (e.g., Glen et al., 2007c). As pointed out by Glen et al. (2007c), the presence of ocean island basalt magmatism in the southern Thomson may actually reflect accretion onto an ‘east-west convergent margin’; Gray and Foster (2004) notably invoked a similar tectonic scenario for the southern Thomson Orogen, which they appear to link to the Larapinta seaway and younger structural events in central Australia (Fig. 22). Glen et al. (2007c) suggested that any collision probably predated the Late Devonian, as sedimentary rocks of this age are found in both the Thomson and Lachlan orogens. The possibility of such a scenario raises questions about the Thomson Orogen as a whole, and how it relates to northern Australia, especially the ‘Tasman Line’ and the geological interpretation of the latter.

Tectonic summary of eastern Australia by time period

2.1. Late Neoproterozoic to Early Cambrian

Rodinian break-up – pre-Delamerian Orogeny: ca. 600 to ca. 520 Ma

The Late Neoproterozoic (ca. 600 Ma) to mid Cambrian geological history of southeastern Australia records episodic glacial and marine sedimentation, thought to be related to global glaciation (e.g., Hoffman and Schrag, 2002), as well as a cycle of continental rifting and ocean opening, related to the breakup of Rodinia, with ensuing formation of passive margins and initiation of the Pacific Ocean (e.g., Li and Powell, 2001; Cawood, 2005; Li et al., 2008; Figs 26, 27). The latter continued in eastern Australia until it was effectively ended by subduction (starting at least by ca. 515 Ma in the southern Delamerian; Foden et al., 2006) and arc-continent collision, ca. 510-505 Ma (Crawford and Berry, 1988, 1992), related to the Delamerian Orogeny. Glen (2005) called this interval the Delamerian Cycle, and suggested that it lasted more than 300 Ma, from ca. 830-780 Ma (see also Li et al., 2008). Most of this period falls outside the time range of this report and is not covered here (see Drexel and Preiss, 1995; Calver and Walter, 2000; Crawford et al., 2003a; Glen, 2005; and references therein for more information).

Marine sedimentary successions of this age occur throughout eastern and central Australia (e.g., Li and Powell, 2001; Fig. 26). In southern Australia, these consist of Neoproterozoic successions such as the ca. 700 Ma Sturtian and ca. 600-580 Marinoan glacial successions of South Australia (e.g., Walter et al., 2000) and similar rocks in Tasmania (Calver and Walter, 2000; Calver et al., 2004) that are suggested to be related to 'snowball earth' and subsequent deglaciation events (e.g., Hoffman and Schrag, 2002). Also present are widely distributed marine sediments which contain widespread evidence outlined below for continental breakup related to Rodinian rifting (e.g., Cawood, 2005; Crawford et al., 2003a). This evidence is best preserved (Figs 26, 27) in rocks ca. 600 Ma in age and younger (to 500 Ma), in the Delamerian Orogen - western Tasmania and King Island (e.g., Calver and Walter, 2000; Calver et al., 2004; Meffre et al., 2004), South Australia (e.g., Drexel and Preiss, 1995; Foden et al., 2001), western Victoria (VandenBerg et al., 2000; Crawford et al., 2003b), the Koonenberry region, western New South Wales (e.g., Crawford et al., 1997; Gilmore et al., 2007), in north Queensland - the Georgetown region (Fergusson et al., 2007a), Charters Towers region (Hutton et al., 1997; Fergusson et al., 2001; 2007c), and in the Thomson Orogen - in the Anakie Inlier (Withnall et al., 1995). As summarised by Crawford et al. (1997; 2003a, b) and Fergusson et al. (2007a, 2007c), many rocks of this age contain alkaline and/or tholeiitic assemblages consistent with rift tectonics and a passive margin and mantle-plume magmatism. Crawford et al. (2003a) suggested rifting was oriented largely northwest-southeast to explain the distribution of rift volcanism at this time (Fig. 27). Palaeogeographic reconstructions of Rodinia often suggest that break-up occurred early and that rifting was well offshore of Australia by 600 Ma, (e.g., Li and Powell, 2001; Li et al., 2008). The abundance of 600-570 Ma rift-related magmatism in eastern Australia would appear to indicate, as suggested by Crawford et al. (2003a), that actual break-up may have begun ca. 600 Ma.

Detrital zircon ages obtained by Fergusson et al. (2007c) in the Thomson Orogen and southern north Queensland Orogen show that the late Neoproterozoic (ca. 600 Ma) rocks (e.g., Cape River Metamorphics, lower Argentine Metamorphics (Charters Towers region) and Bathampton Metamorphics (Anakie Inlier) generally contain abundant ca. 1200-1000 Ma (Grenville-age) detrital zircons and only minor 1870-1550 Ma zircons, indicating at best only minor input from (present-day nearby) cratonic regions such as Mount Isa and Georgetown. Fergusson et al. (2007c) suggested the ca. 1200 Ma zircons were possibly derived from an extension of the Late Mesoproterozoic (1200–1050 Ma) orogenic belt represented by the Musgrave Inlier (central Australia) 1500 km to the west. Maidment et al. (2007) have suggested that similar zircon populations in the Amadeus Basin, central Australia, reflected uplift and erosion of the Musgrave Inlier during the Petermann Orogeny.

Late Neoproterozoic and Early (to Late) Cambrian rocks also occur within the Lachlan and New England orogens (Fig. 26). These fall into two main types. The first type consists of mafic and ultramafic Cambrian (and older?) igneous rocks, preserved as remnants along major faults, in the Victorian part of the Lachlan (e.g., VandenBerg et al., 2000; Spaggiari et al., 2004; Fig. 26). These include an ultramafic and a tholeiitic-boninitic association (e.g., Crawford and Keays, 1987; Crawford et al., 1984, 2003b; VandenBerg et al., 2000), that are typically interpreted as having formed in a suprasubduction zone environment (e.g., Crawford and Keays, 1987; Crawford et al., 1984, 2003a, b; Spaggiari et al., 2003, 2004). In the Stawell Zone, the mafic volcanics (Magdala Volcanics) have a backarc signature and are underlain by continental-derived turbiditic sediments (Crawford et al., 2003b; Squire et al., 2006; Fig. 27). These authors interpreted the Stawell succession to represent a distal backarc environment, related to a west-dipping subduction zone to the east. This is also consistent with the observation that at least some of these mafic-ultramafic successions, for example, in the Bendigo and Tabberabbera zones, appear to form the basement to those zones (e.g., Spaggiari et al., 2003, 2004; Korsch et al., 2008), that is, floored by oceanic crust as suggested by Gray, Foster and co-workers (e.g., Gray and Foster, 2004). These are overlain, conformably in places, by Cambrian, deep marine, often pelagic sedimentation (VandenBerg et al., 2000; Spaggiari et al., 2003). Examples, such as those in the Bendigo Zone and further east, were not affected by the Delamerian Orogeny (Spaggiari et al., 2003, 2004). Mafic-ultramafic successions in the west, including those in the western Stawell Zone, were deformed during the Delamerian Orogeny (e.g., Miller et al., 2006). The New England Orogen also contains similar mafic-ultramafic remnants. These are Neoproterozoic to Cambrian tectonic blocks, largely of oceanic fragments, including island arc-related remnants (Fig. 26). They occur in the southern NEO, as accreted blocks along the Peel-Manning Fault System (e.g., Offler and Shaw, 2006), and in the northern NEO, as represented by the ca. 565 Ma Princhester and related ophiolites (Bruce et al., 2000; Murray and Blake, 2005). In both the Lachlan and New England orogens, these rock associations provide records of subduction and other oceanic environments outboard of continental Australia in the Neoproterozoic and Cambrian (Fig. 27). The initiation of Early Palaeozoic subduction in the southern New England Orogen is recorded by the formation of suprasubduction zone ophiolites (~530 Ma; Aitchison and Ireland, 1995; Fanning et al., 2002; Sano et al., 2004), as well as blocks of Late Neoproterozoic eclogite and intrusive rocks (~530 Ma ages; Aitchison et al., 1992a; Sano et al., 2004; Watanabe et al., 1998; Fanning et al., 2002), and Cambrian magmatic island arc development (Cawood and Leitch, 1985). Whereas contacts between exposed intraoceanic elements are faulted, their character and distribution suggest development in an east-facing arc with the Cambrian ophiolite separating, and underlying, the western arc-flanking basin from the eastern accretionary prism (Cawood and Leitch 1985; Holcombe et al., 1997a, b; Jenkins et al., 2002). This record of Cambrian subduction is slightly earlier than boninitic magmatism recorded in Tasmania (514±5 Ma; Black et al., 1997) and in Victoria (519-514 Ma; VandenBerg et al., 2000). Rocks at Port Macquarie also contain a fragmentary late Neoproterozoic to early Palaeozoic history that includes subduction and associated metamorphism under high P- low T conditions at ~536 Ma (Watanabe et al., 1998; Och et al., 2007). These ages suggest subduction had begun by at least 530 Ma (e.g., Li and Powell, 2001), and possibly earlier (e.g., Gray and Foster, 2004).

Also present within the Lachlan Orogen are interpreted Delamerian-aged and older rocks of the Selwyn Block (VandenBerg et al., 2000). The Selwyn Block hypothesis proposes the presence of older continental basement beneath the Melbourne Zone, which has been linked to western Tasmania (Cayley et al., 2002; Fig. 26). Although some authors have suggested oceanic crust, only, as basement for the Lachlan Orogen (e.g., Gray, 1997; Gray and Foster, 1997, 1998; Figs 24, 25), the presence of the Selwyn Block appears to be confirmed by the recent Victorian seismic acquisition (e.g., Korsch et al., 2008; Cayley et al., in prep). The latter clearly shows the Melbourne Zone to be underlain by something distinct from zones to the west. The Selwyn Block contains Cambrian calcalkaline volcanics (e.g., VandenBerg et al., 2000; Spaggiari et al., 2003), which have many similarities to, and have been correlated with, the Mount Read Volcanics (Crawford et al., 2003a, b). Crawford et al. (2003a) have suggested that the Selwyn Block represented part of Australia rifted off during the 600 Ma breakup event (Fig. 27). The Anakie Inlier may have formed in a manner similar to the Selwyn Block; Fergusson et al. (2001) have suggested it may have been related to splitting or rifting of a younger microcontinent from the Gondwanan margin. The sparse available data for the Thomson Orogen make this difficult to prove or disprove.

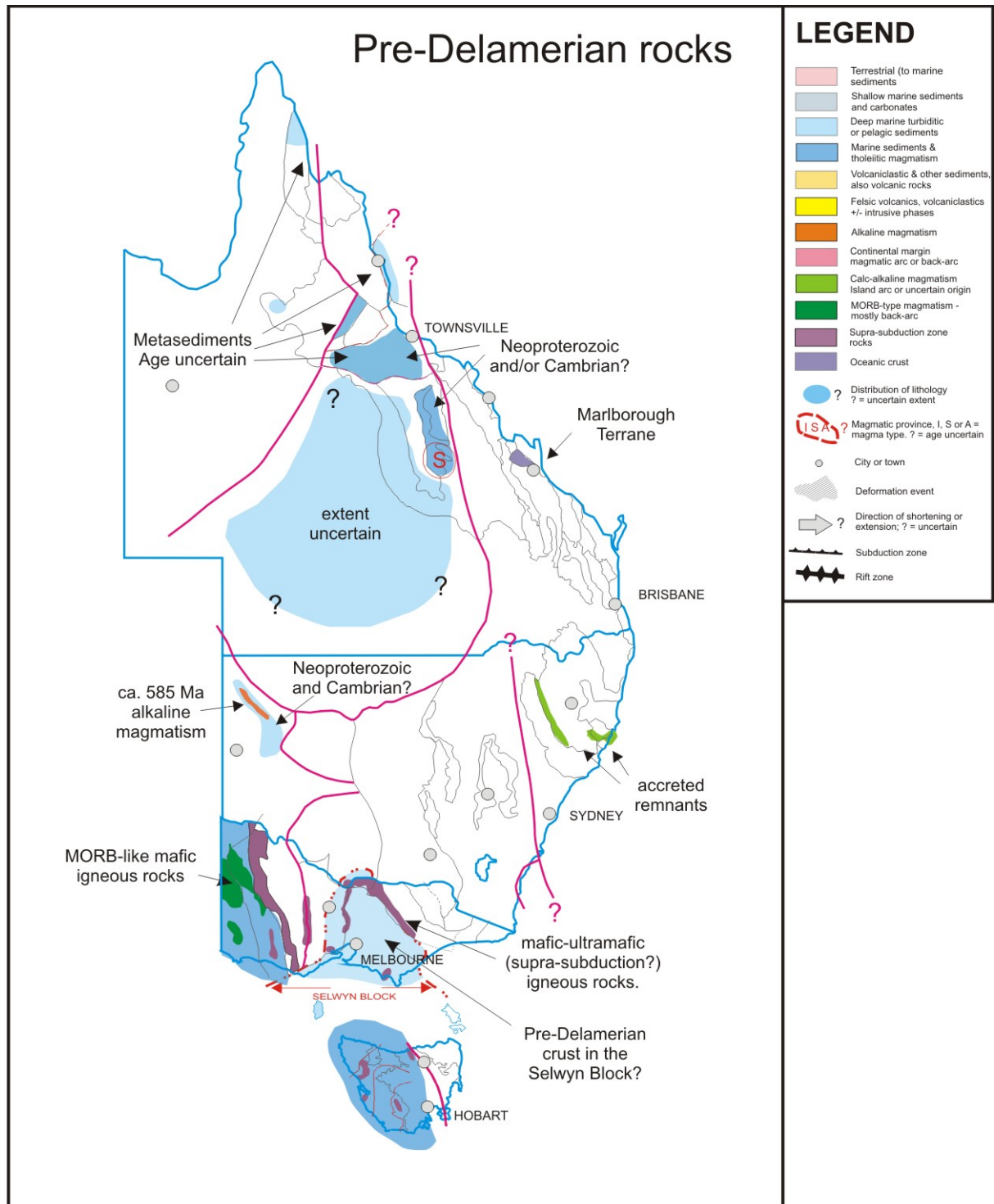


Figure 26. General distribution of Neoproterozoic to mid Cambrian (ca. 600 Ma to 515 Ma), pre-Delamerian rocks in eastern Australia. Refer to text for discussion.

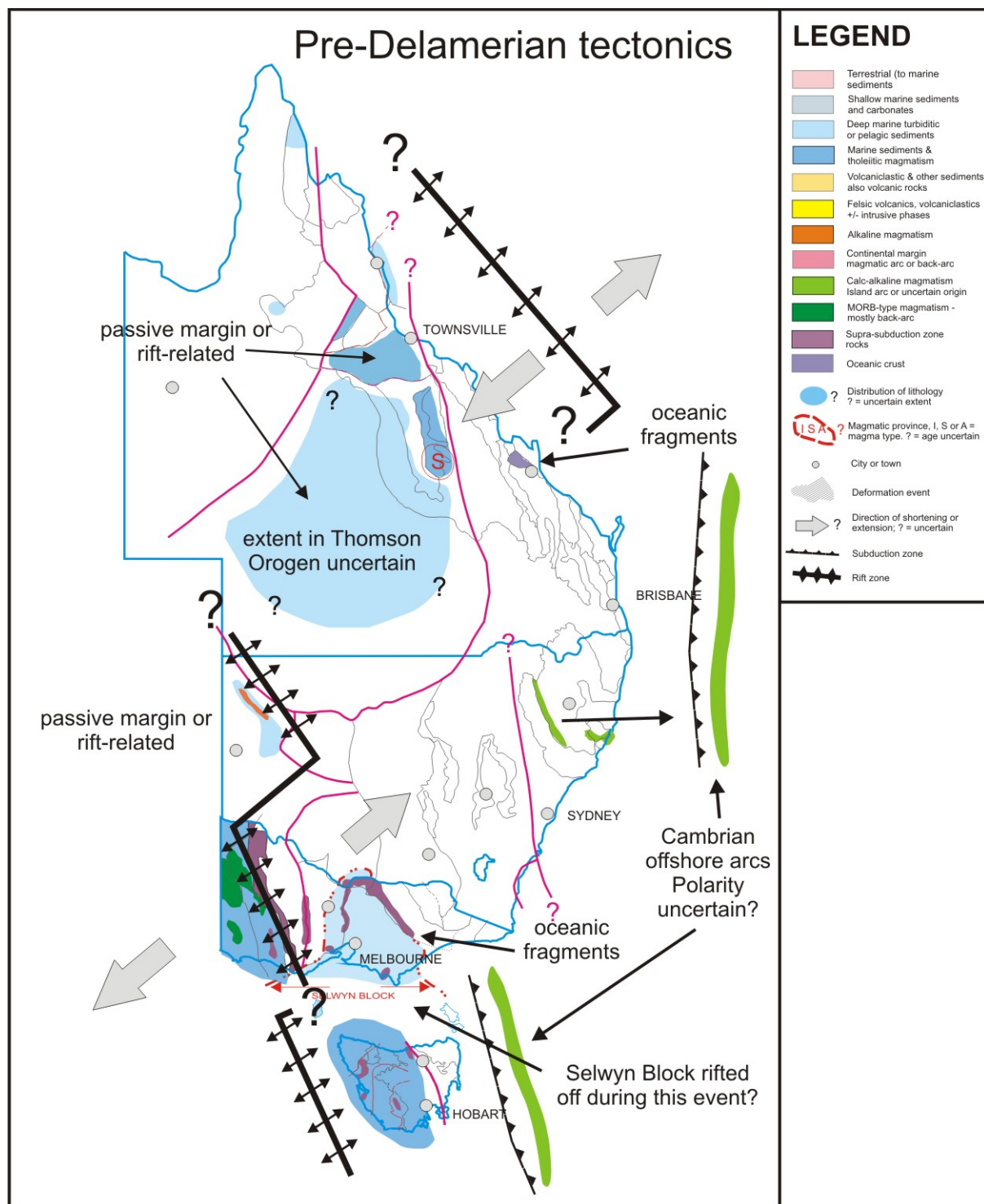


Figure 27. Interpreted tectonic environment of eastern Australia for the pre-Delamerian period (Neoproterozoic to mid Cambrian - ca. 600 to Ma 515 Ma). Refer to text for discussion of tectonic interpretation. Location of the Melbourne Zone and Tasmania in this time period is uncertain.

2.2. Early to Middle Cambrian:

Delamerian Orogeny: ca. 520 Ma to ca. 490 Ma

The breakup of Rodinia and associated extension of southeastern Australia between the Late Neoproterozoic (ca. 600 Ma) and the Early Cambrian was halted with the onset of subduction and accompanying contractional orogenesis – called the Delamerian Orogeny. In South Australia, western Victoria and Tasmania, the Delamerian Orogeny (Tyennan Orogeny in Tasmania) commenced at ca. 515 Ma (e.g., Foden et al., 2006; Seymour and Calver, 1995; Figs 28, 29). In South Australia, the Delamerian Orogeny was long-lived, from ca. 515 to 490 Ma (e.g., Drexel and Preiss, 1995; Foden et al., 2006). VandenBerg et al. (2000), amongst others, suggested that there may have been two stages, ca. 515 and ca. 490 Ma, and indicated that, in Victoria, the first phase is evident in the Glenelg Zone, the second only in the Grampians-Stavely Zone. Miller et al. (2006) showed that the Delamerian Orogeny also affects the western part of the Stawell Zone, based on metamorphic ages of ca. 500-490 Ma. Western Tasmania also records at least two discrete deformational events, separated by a significant extension event, and VandenBerg et al. (2000) showed that, in many respects, western Victoria and western Tasmania share similar geological histories. The Delamerian Orogeny in both western Tasmania and western Victoria was apparently triggered by arc-continent collision around 515-510 Ma (Berry and Crawford, 1992; VandenBerg et al., 2000; Crawford et al., 2003a), possibly by an east-dipping subduction zone (e.g., Crawford et al., 2003a; though possibly flipping to west-dipping after collision). In both areas, collision was accompanied by the accretion of Cambrian forearc boninitic crust – the Tasmanian mafic-ultramafic complex in Tasmania (Crawford and Berry, 1992; Crawford et al., 2003a), and the Dimboola Igneous Complex in western Victoria (VandenBerg et al., 2000; Crawford et al., 2003b; Fig. 28). High temperature, low pressure, metamorphism and metamorphic complexes were developed in both regions, with syntectonic I- and S-type granites emplaced in the Glenelg River Metamorphic Complex in Western Victoria (VandenBerg et al., 2000; Crawford et al., 2003b). Both regions underwent subsequent post-collisional extension (possibly in a backarc environment), leading to emplacement of calcalkaline volcanics – Mount Read Volcanics, correlatives, and intrusives, in Tasmania (Crawford and Berry, 1992; Crawford et al., 2003a), and the Mount Stavely Volcanic Complex in Victoria (Crawford et al., 1996, 2003b; VandenBerg et al., 2000; Fig. 28). This was followed by renewed deformation at ca. 500 Ma to 490 Ma.

Evidence for the Delamerian Orogeny is also found in New South Wales and Queensland. Although broadly similar ages to South Australia and Victoria have been recognised, deformation appears to have been shorter-lived farther to the north (Foden et al., 2006; Black, 2007; Fergusson et al., 2007a, b). The Delamerian Orogeny deformed and metamorphosed rocks of the Anakie Inlier, Charters Towers region (see below), Koonenberry Belt and central Thomson Orogen, and mostly resulted in downwarping of the Warburton Basin (Murray and Kirkegaard, 1978; Fig. 30). In this region, the Delamerian Orogeny is best recorded in basement rocks of the Anakie and Koonenberry regions. Based on comparative evidence from the Anakie Inlier, Draper (2006) suggested that deformation of subsurface metasedimentary rocks in the eastern Thomson Orogen was also likely to have occurred during the Delamerian Orogeny. The contractional event was predominantly east-west in the Thomson Orogen, and northwest-southeast in the Koonenberry Belt (Gilmore et al., 2007). The Delamerian Orogeny is poorly recorded in the Warburton Basin in general (e.g., Murray and Kirkegaard, 1978) though uplift and erosion of the basin was apparently coincident with the orogeny (recognised as the Mootwingee Movement; Gravestock and Gatehouse, 1995).

Delamerian deformation was accompanied by syn- to postorogenic calcalkaline magmatism in the Anakie Inlier and Koonenberry Belt (Withnall et al., 1995; Crawford et al., 1997; Gilmore et al., 2007), and in the Warburton Basin (Gatehouse, 2006). As suggested by Gatehouse (1986), the calcalkaline volcanism may relate to an arc present at this time. Although the Bourke-Louth regions are adjacent to the Koonenberry, rocks of this age do not appear to have been recorded in the former. The possible presence of an arc in the Warburton-Koonenberry region at this time, well west of the Anakie Inlier, is problematical.

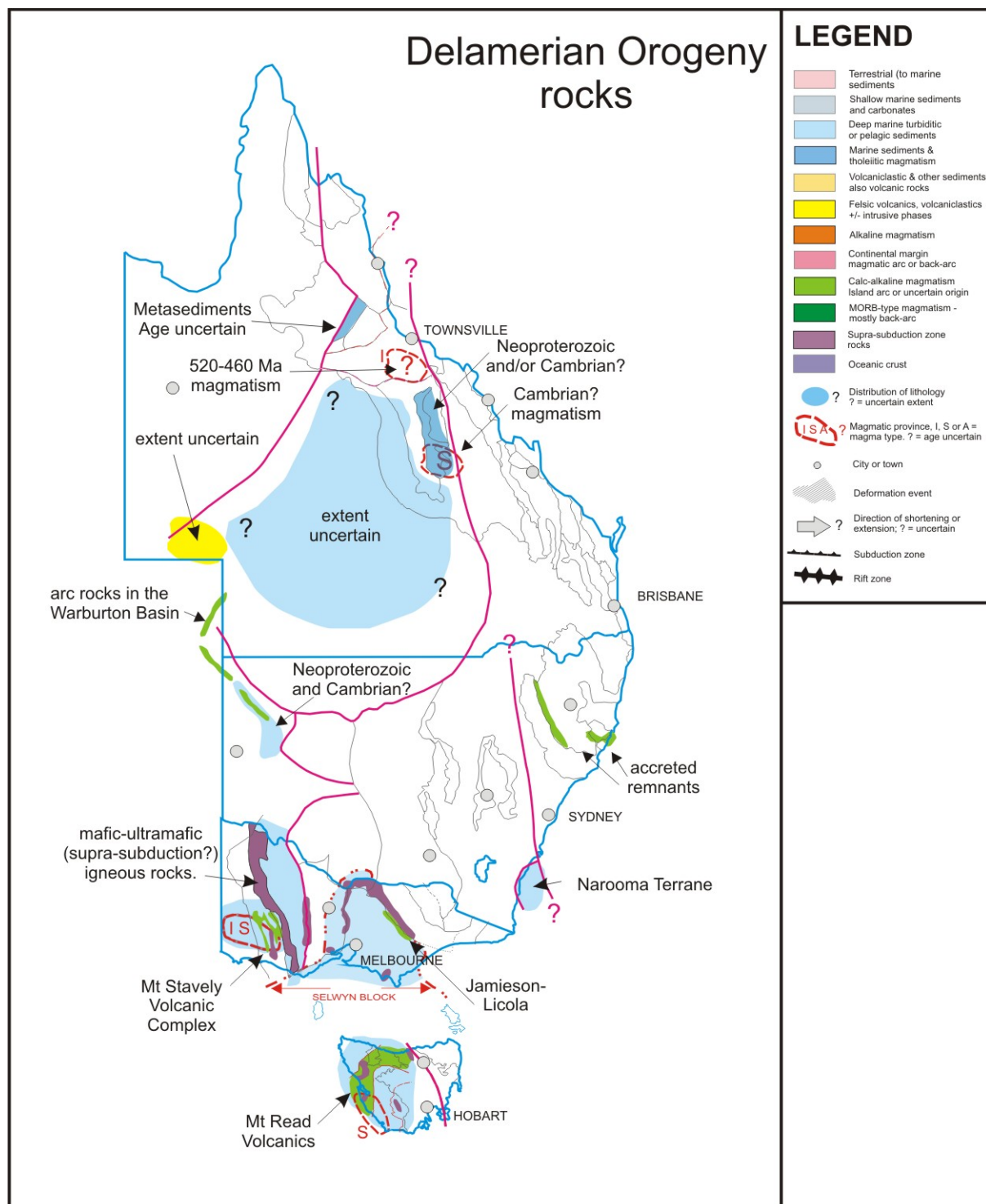


Figure 28. General distribution of mid to Late Cambrian (ca. 520 Ma to ca. 490 Ma), Delamerian cycle rocks in eastern Australia. Refer to text for discussion.

Withnall et al. (1995) suggested this may have either reflected a very wide Delamerian Orogen or subsequent (post-Delamerian) extension and rifting of the Anakie Inlier eastwards. The latter is, however, probably not consistent with the recent discovery of flat-lying Ordovician Volcanic rocks overlying deformed and metamorphosed metasedimentary rocks in the central Thomson Orogen (Draper, 2006).

The Delamerian Orogeny is poorly represented in northern Queensland, partly reflecting the geochronological uncertainty of many of the units. The best evidence for this orogeny appears to be within the Greenvale Subprovince (ca. 520-510 Ma; Nishiya et al., 2003) and in the Charters Towers region (ca. 495 Ma; Fergusson et al., 2007a). Potential Delamerian deformational events may occur in the Georgetown and Coen regions, but geochronological data are missing. A number of deformations that could be interpreted as Delamerian have been shown, at least partly, to represent post-Delamerian extension (Fergusson et al., 2007a, b). There is little definitive evidence for the existence of an arc environment at this time in north Queensland. Possibly the best indirect indication of such an arc are the post-Delamerian, latest Cambrian-Early Ordovician, volcanic successions in the Georgetown and Charters Towers regions (Henderson, 1986; Withnall et al., 1991; Stolz, 1994), usually interpreted as backarc settings.

The Selwyn Block, Victoria, contains Cambrian calcalkaline volcanics (Jamieson-Licola, e.g., VandenBerg et al., 2000; Spaggiari et al., 2003; Fig. 28, 30), which have similarities to, and have been correlated with, the Mount Read Volcanics in Tasmania (Crawford et al., 2003a, b). Crawford et al. (2003b) suggested they may represent 'along strike continuations' of the Mount Read Volcanics. In the oceanic crust basement model, these calcalkaline rocks are interpreted as island arc volcanics, e.g., Jamieson Island Arc (Gray and Foster, 2004; Spaggiari et al., 2003; Fig. 25). Other evidence for island arc terranes of this age are present in the New England Orogen (e.g., Offler and Shaw, 2006), as well as in New Zealand, e.g., the ca. 515 Ma Takaka Island Arc (e.g., Munker and Crawford, 2000). The southern New England Orogen in this period comprises Cambrian tectonic blocks, largely of oceanic fragments, including island arc-related remnants, such as suprasubduction zone ophiolites (~530 Ma; Aitchison and Ireland, 1995; Fanning et al., 2002; Sano et al., 2004; Fig. 30) and Cambrian magmatic arc rocks (Cawood and Leitch, 1985), which provide a record of continuing subduction environments offshore of continental Australia in the pre-Delamerian and Delamerian, that is, (latest Neoproterozoic-) Cambrian. These remnants have been interpreted to suggest an east-facing arc (e.g., Cawood and Leitch 1985). Rocks at Port Macquarie also contain late Neoproterozoic to early Palaeozoic remnants that indicate subduction and associated metamorphism under high P low T conditions at ~536 Ma (Watanabe et al., 1998; Och et al., 2007). Cambrian magmatic arc development in New England is also recorded by Middle to early Late Cambrian volcanoclastic rocks, apparently derived from a low-K intra-oceanic island arc (Cawood and Leitch, 1985). These rocks - Murrawong Creek and Pipeclay Creek Formations - occur in the Gamilaroi Terrane immediately west of the Peel-Manning Fault. These oceanic remnants in the Lachlan and New England Orogens provide important tectonic constraints. They represent fragments of oceanic and island arc crust that were accreted during the Delamerian Orogeny and later, and are consistent with eastern Australia not only facing the Palaeopacific Ocean since the late Neoproterozoic and earliest Palaeozoic but also consistent with an overall arc environment for much (all?) of this time (e.g., Crawford et al., 2003a; Gray and Fergusson, 2004; Collins and Richards, 2008; Li et al., 2008; Fig. 22). They also indicate that parts of the now contiguous orogens were probably significantly separated in the Early Palaeozoic, as suggested by many authors (e.g., VandenBerg et al., 2000; Cayley et al., 2002; Gray and Foster, 2004; Fig. 30).

The polarity of subduction in the Delamerian is uncertain, and both east-dipping (e.g., VandenBerg et al., 2000; Munker and Crawford, 2000; Crawford et al., 2003a) and/or west-dipping (e.g., Gray and Foster, 2004; Foden et al., 2006; Figs 22, 30) subduction has been invoked. It is also possible that subduction polarity has switched (e.g., Munker and Crawford, 2000), such that it may have been east-dipping pre-collision and west-dipping post-collision, similar to the model of Murray (2008) for the Devonian Calliope Island Arc in Queensland. Perhaps the best evidence for post-collision polarity is provided by the backarc volcanic rocks in the Stawell Zone (see previous section), which Squire et al. (2006) interpreted to represent a distal backarc environment, related to a west-dipping subduction zone to the east (Fig. 30).

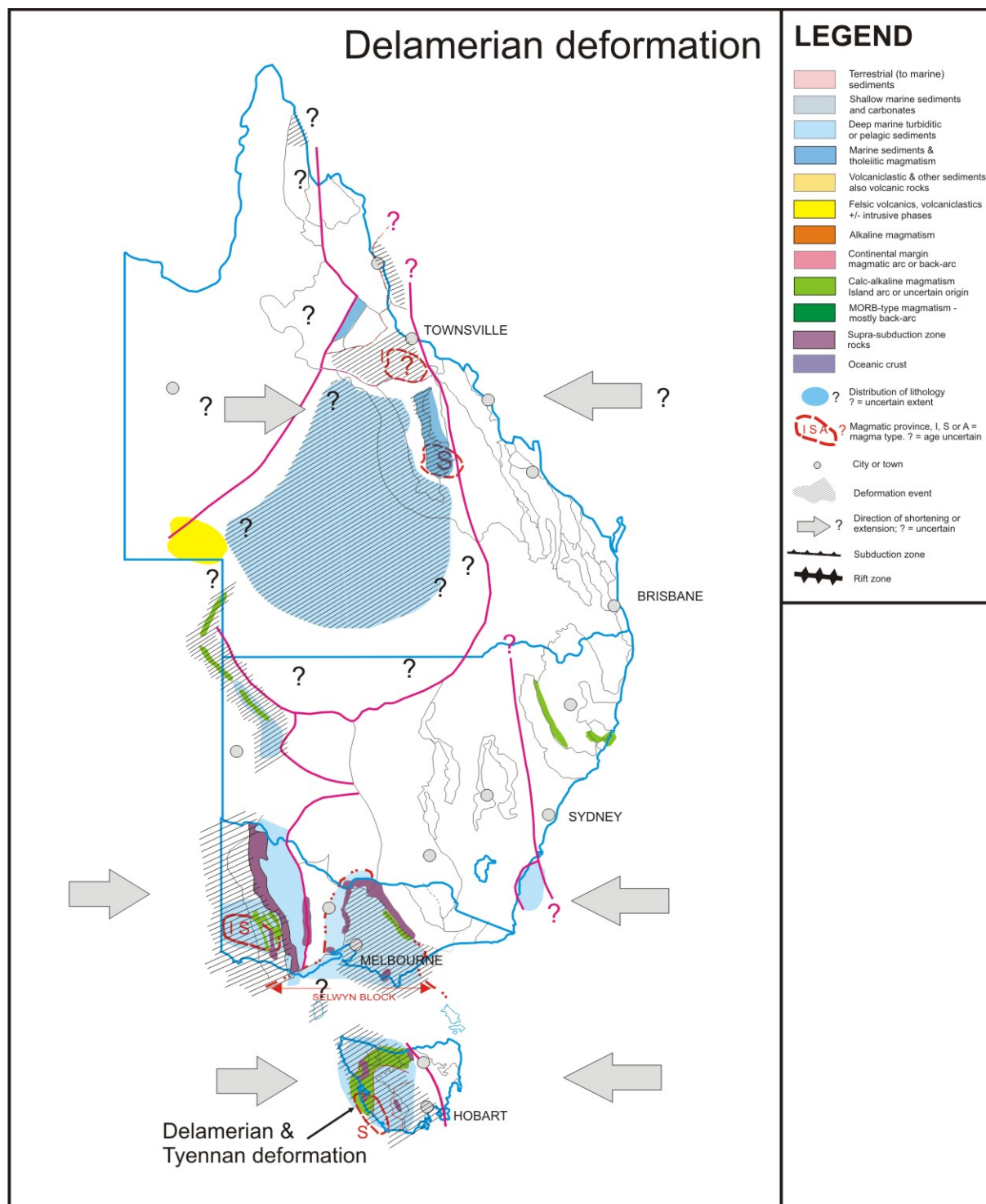


Figure 29. Generalised (approximate) distribution of deformation within the Delamerian Orogeny (ca. 520-490 Ma) as defined by outcrop and drill hole information (Thomson Orogen), except for the Selwyn Block. Extent of deformation in the latter has been extrapolated from minor outcrops, e.g., Waratah Bay (e.g., VandenBerg et al., 2000). For all areas actual extent of deformation may be greater and more continuous. The intensity of deformation is variable within the areas indicated. See text for data sources.

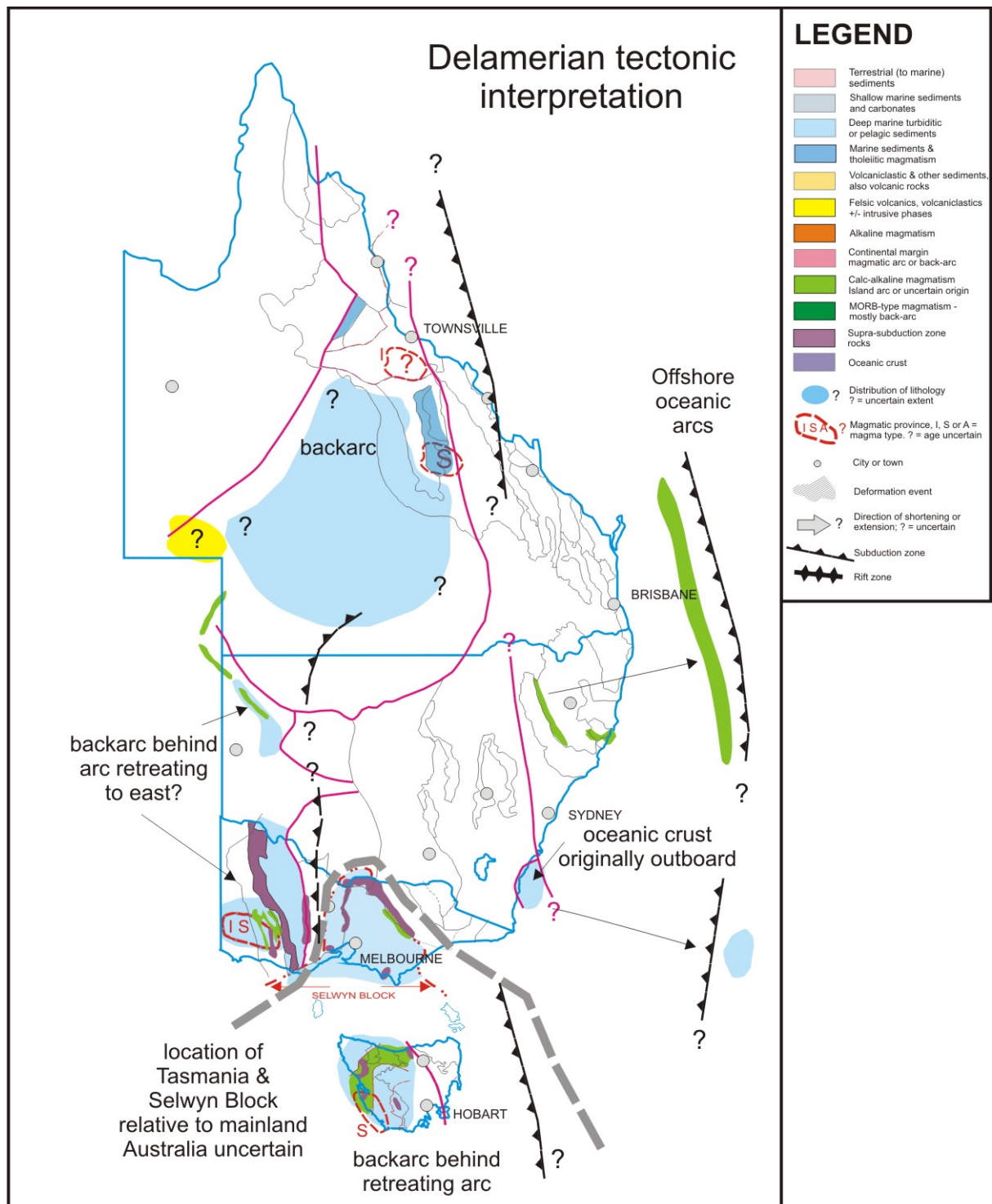


Figure 30. Interpreted tectonic model for the Delamerian Orogeny of eastern Australia (ca. 520-490 Ma). Refer to text for detailed discussion. Location of the Melbourne Zone and Tasmania in this time period is uncertain.

Similarly, calcalkaline volcanics in western Tasmania, the Melbourne Zone, the Kooneberry region and the Warburton Basin (Figs 28, 30) all appear consistent with a west-dipping arc. As noted by Squire et al. (2006), this scenario also explains the observed temporal sequence (forearc to backarc) evident in many of the mafic-ultramafic complexes (seafloor remnants) preserved in both the Delamerian and Lachlan orogens (Crawford et al., 2003b). Following the Delamerian Orogeny, the arc appears to have shifted well to the east, for example, as prepresented by the interpreted Ordovician Macquarie Arc (Crawford et al., 2007a), resulting in post-Delamerian extension, sedimentation and emplacement of post-tectonic granites (e.g., VandenBerg et al., 2000; Foden et al., 2006).

2.3. Late Cambrian to earliest Silurian

Post-Delamerian to Benambran Orogeny: ca. 490- ca. 430 Ma

Eastern Australia through the Benambran Cycle (post-Delamerian orogeny to Benambran Orogeny) is dominated by two contrasting rock packages:

- deep water quartz-rich turbidites of cratonic provenance and associated pelagic sediments, commonly interpreted as a backarc and/or passive margin environment.
- calcalkaline magmatism and volcanoclastics and marine sediments with common carbonates, commonly interpreted as having formed in oceanic arcs, and/or backarc environments.

These contrasting rock packages are best exemplified in the Lachlan Orogen. From the Late Cambrian to the end of the Ordovician-Early Silurian, most of the Lachlan Orogen (New South Wales, Victoria and Tasmania) was the site of deep marine sedimentation (Fig. 31). These sediments consist of quartz-rich turbiditic successions (Western, Central and Eastern Lachlan) and late Middle to Late Ordovician black shale-dominated sediments (Central and Eastern Lachlan). In a number of regions, for example, the Bendigo and Tabberabbera zones (e.g., VandenBerg et al., 2000; Fergusson and VandenBerg, 2003; Spaggiari et al., 2003), this sedimentation appears to be conformable upon mafic and ultramafic rocks interpreted as oceanic crust (see previous section). Remnants of mafic volcanics, chert, serpentinites and ultramafic rocks interpreted as oceanic crust also occur within New South Wales (e.g., Warren et al., 1995) and are also interpreted to probably represent basement to the turbidite successions. In most areas of the Lachlan Orogen, this deep water sedimentation ended with the Benambran Orogeny. Sedimentation, however, continued in both the Melbourne Zone and northeastern Tasmania, and both these regions show no evidence for the Benambran Orogeny (e.g., Fergusson and VandenBerg, 2003; Seymour and Calver, 1995; Fig. 32).

Contemporaneous with quartz-rich turbiditic sedimentation was the Early Ordovician to earliest Silurian Macquarie Arc, commonly interpreted to have formed in an intra-oceanic arc setting (e.g., Crawford et al., 2007a), though Wyborn (1992), for example, suggests an alternate, non-arc, setting. The remnants of the arc, which include calcalkaline and shoshonitic volcanic rocks, intrusions and volcanoclastic and carbonate-rich successions, are preserved as four elongate remnants, almost totally located in New South Wales though extending into northern Victoria (Fergusson and VandenBerg, 2003; Fig. 31). Recent detailed work (Crawford et al., 2007a and companion papers) suggests that the Macquarie Arc was built up over four successive phases of growth, within two distinct (east and west) provinces, which may not have been together until accretion (Percival and Glen, 2007). Other possible exotic oceanic rocks, interpreted to have accreted to Australia during the Benambran orogeny, are the deep marine sediments and underlying mafic volcanics of the Narooma Terrane (Glen et al., 2004; Fig. 31). Miller and Gray (1997), however, suggested these rocks do not represent an exotic terrane, but formed within an accretionary wedge.

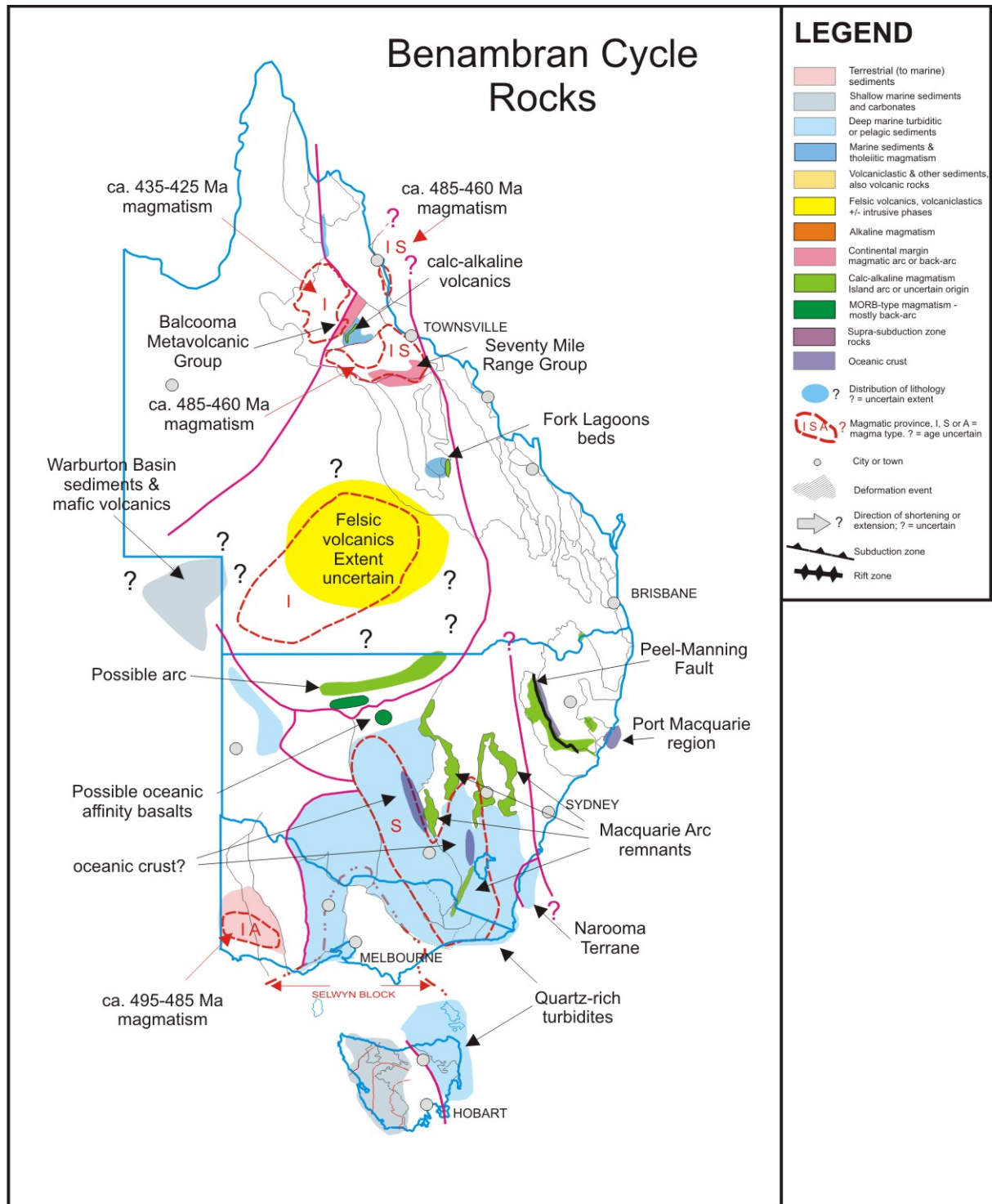


Figure 31. General distribution of latest Cambrian to Early Silurian Benambran (ca. 490 Ma – ca. 430 Ma) cycle rocks in eastern Australia. Refer to text for discussion.

Most workers indicate that one or more submarine fans were the sites for the Ordovician deep marine sedimentation (e.g., Fergusson and VandenBerg, 2003; Gray and Foster, 2004; Glen, 2005; Glen et al., 2007b). This may have reflected uplift of the Delamerian Orogen in the Early Ordovician (e.g., VandenBerg et al., 2000; Fergusson and VandenBerg, 2003), although there is uncertainty regarding the possible relative positions of the respective parts of the Lachlan Orogen at this time. The major change in sedimentation recorded by the switch to black shale-dominated pelagic sediments in the late Middle Ordovician and their localisation largely to the Central and Eastern Lachlan, is in part contemporaneous with early Benambran deformation (ca. 455 Ma) recorded in the Western Lachlan (VandenBerg et al., 2000; Gray et al., 2003; Miller et al., 2006).

Importantly, these turbiditic sediments apparently provide no evidence for volcanic detritus, and it would appear, as suggested by numerous workers (e.g., Gray and Foster, 2004; Meffre et al., 2007), that the contemporaneous Macquarie Arc was disconnected from this deep marine sedimentation, perhaps by hundreds of kilometres (Meffre et al., 2007). The tectonic environment for the quartz-rich sediments is commonly interpreted as a passive margin environment, often in a backarc position, well behind the Macquarie Arc (e.g., Fergusson and VandenBerg, 2003; Gray and Foster, 2004; Glen, 2005; Glen et al., 2007b; Fig 22). The quartz-rich sediments and the Macquarie Arc are thought to have been juxtaposed when the arc accreted to eastern Australia, in the Early Silurian as part of the Benambran event (Glen et al., 2007b; Meffre et al., 2007). In the non-arc model of Wyborn (1992) there is no requirement for accretion, though the apparent lack of interfingering of quartz-rich and volcanic sediments is problematical.

Although largely dismembered, the north Queensland region (e.g., Henderson, 1987), like the Lachlan Orogen, consists of both quartz-rich sediments and calcalkaline volcanism (Fig. 31). The region consists of Ordovician deep water (turbiditic), dominantly quartz-rich sediments, preserved within fault-bounded remnants east of, and probably derived from, the outcropping Mesoproterozoic basement in the Georgetown and Coen regions (e.g., Withnall and Lang, 1993; Garrad and Bultitude, 1999; Fig. 31). The sediment successions in north Queensland also contain interlayered tholeiitic magmatism (Withnall and Lang, 1993; Bultitude et al., 1997; Withnall et al., 1997a), consistent with an extensional environment. Contemporaneous calcalkaline magmatism is preserved as Early to Middle Ordovician volcanic- or volcanoclastic-dominated successions (e.g., Seventy Mile Range Group, Balcooma Metavolcanic Group - Henderson, 1986; Withnall et al., 1991; Stolz, 1995; Fergusson et al., 2007a) and Early and Late Ordovician volcanic and carbonate-dominated sequences, for example, in the Broken River Province (Withnall and Lang, 1993; Fig. 31). Many authors have suggested backarc, continental-margin arc or island-arc affinities for the sediments and calcalkaline successions (e.g., Withnall et al., 1991, 1997b; Henderson, 1986; Stolz, 1994), suggesting an environment not dissimilar to Lachlan Orogen rocks of the same age (e.g., Gray and Foster, 2004; Glen, 2005). Unlike the Lachlan Orogen, however, units in north Queensland locally contain both quartz-rich marine sediments and calcalkaline volcanics (e.g., the Judea Formation; Withnall and Lang, 1993), as well as volcanic clasts in conglomerate which appear to correlate with known calcalkaline volcanic units in the region (Garrad and Bultitude, 1999). These features suggest proximity between arc-related volcanism and cratonic-derived sedimentation, and support suggestions that the quartz-rich sediments were deposited within a back-arc environment. A similar scenario is recorded within the Anakie Inlier (eastern Thomson Orogen) which contains Late Ordovician marine sediments, including carbonates, and associated mafic to intermediate volcanic rocks (Fork Lagoons Beds, Withnall et al., 1985; Fig. 31). The sediments appear to be derived from cratonic and volcanic provenances (Fergusson et al., 2007c) and the volcanic rocks have geochemistry consistent with either arc or backarc environments (Withnall et al., 1995). Calcalkaline volcanics, interpreted as arc-related, are also recorded in the southern Thomson Orogen, in New South Wales (e.g., Burton et al., 2008). According to Watkins (2007) and Burton et al. (2008), these appear to be chemically similar to (nearby) Macquarie Arc magmatism (Fig. 31). Notably, contemporaneous within-plate magmatism is also recorded, which Watkins (2007; Figs 31, 33) placed in a backarc environment.

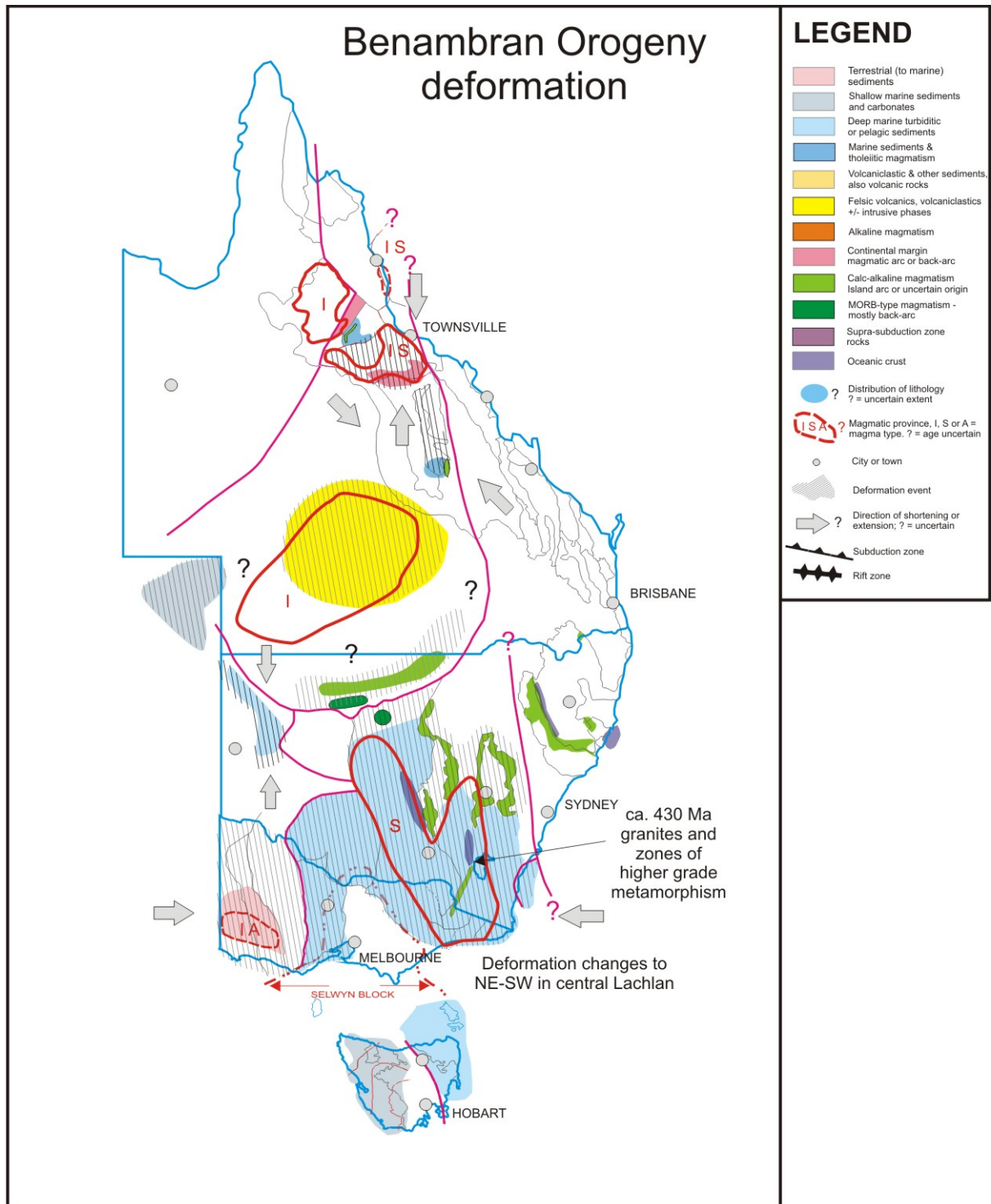


Figure 32. Generalised (approximate) distribution of deformation (cross-hatching) within the Benambran Orogeny (ca. 455 Ma – ca. 430 Ma) as defined by outcrop, drill core and extrapolation – actual extent of deformation may be greater and more continuous. The intensity of deformation is variable within the areas indicated. Also highlighted in continuous red lines are regions with syn-tectonic, ca. 430 Ma granites. See text for data sources. Refer to [Figure 31](#) for additional geological explanations.

Early to Middle Ordovician felsic (rhyolitic) volcanism and associated granites, occur within the central Thomson Orogen. Chemical affinities are unknown. Draper (2006) suggested the magmatism may relate to extension and crustal thinning in the Ordovician. West of the Thomson Orogen, this time period corresponds with the onset of more widespread marine basin sedimentation following the Delamerian Orogeny and intracratonic rifting (e.g., in the Warburton Basin; Gatehouse and Cooper, 1986). Progressive marine incursion caused an expansion of basins across central Gondwana as a shallow epiherc sea (Larapintine Sea; Webby, 1978), believed to link the Warburton, Amadeus and Canning Basins (Webby, 1978; Li and Powell, 2001; Gravestock and Gatehouse, 1995; Maidment et al., 2007). This seaway may have also connected basinal environments in the Koonenberry region, as suggested by Webby (1978), although this is unproven. Sedimentation continued until interrupted by the Benambran/Alice Springs (1) Orogeny which caused uplift and associated regression of the Larapintine Sea (Webby, 1978; Gravestock and Gatehouse, 1995). Contemporaneous shallow marine and terrestrial sedimentation also occurred over Delamerian rocks of western Tasmania (Seymour and Calver, 1995), but not western Victoria.

The history of the New England Orogen in the Late Ordovician and Silurian is fragmentary, but probably involved at least some periods of convergent margin activity (Cawood and Leitch, 1985). In the southern NEO, Late Ordovician rocks are recorded both west and east of the Peel-Manning Fault (Fig. 31). To the west of the fault, arc-derived sediments in fault blocks, with blocks of limestone (Cawood, 1976), lie unconformably on Early Ordovician and Cambrian strata (Cawood and Leitch, 1985). Siliciclastic sedimentary rocks and fossiliferous limestones (Hall, 1975) occur within the imbricate zone of the fault system, and include the Trelawney beds and Haedon Formation (Gamilaroi Terrane). East of the fault, Late Ordovician limestone is recorded below the Silverwood Group (Wass and Dennis, 1977). Late Ordovician coral limestone, probably representing partly accreted seamounts, was deposited in an ocean basin environment and is now incorporated into the Woolomin Terrane (435-428 Ma; Hall, 1978). Middle to Late Ordovician rocks of the Watonga Formation at Port Macquarie have been interpreted to have accumulated on an oceanic plate during its passage from spreading ridge to trench (Och et al., 2007).

Widely distributed post-Delamerian intrusive magmatism occurred across northern- and central-eastern Australia in this cycle (Fig. 31). In north Queensland, ca. 490 Ma to ca. 455 Ma (e.g., Hutton et al., 1997), dominantly felsic I-type, and mafic, (mantle-derived) magmatism comprises the Macrossan Province (Bain and Draper, 1997). These magmatic rocks are best represented in the Charters Towers region. Intrusive magmatism of this age has also recently been confirmed for the Thomson Orogen (e.g., Draper, 2006), although the actual extent is uncertain (Murray, 1994). The Ordovician magmatism in north Queensland is temporally and, at least partly spatially, associated with extensional deformation and low-P high-T metamorphism documented by Fergusson et al. (2007a, b) for the volcanic and sedimentary successions in the Greenvale and Charters Towers regions. Fergusson et al. (2007a, b) suggested this extension was related to backarc development. Such a scenario for the southern Charters Towers region, suggests that the Macrossan Province magmatism in the northern part of the region may represent the actual magmatic arc (e.g., Henderson, 1980; Fig. 33). The only recorded older intrusive magmatism in southeastern Australia (apart from the Macquarie Arc) consist of post-tectonic A- and I-type magmatism in the southern Delamerian Orogen, e.g., in the Glenelg Zone, Victoria, and South Australia (Foden et al., 2006; Fig. 31).

Younger, syn- (to post-) Benambran (ca. 430 Ma and younger) intrusive magmatism occurs throughout eastern Australia, and represents the first manifestation of the voluminous Silurian to Devonian magmatism present within the Lachlan, Thomson and north Queensland orogens (e.g., Chappell et al., 1988; Murray, 1994; Bain and Draper, 1997; Fig. 34). This includes the dominantly I-type magmatism in north Queensland (Georgetown and possibly the Charters Towers region; Withnall et al., 1997a; Hutton et al., 1997; Fergusson et al., 2007a; Fig. 31), the Thomson Orogen (ca. 430 Ma; Draper, 2006), and largely S-type magmatism of early Silurian age in the Central and Eastern Lachlan Orogen (e.g., Collins and Hobbs, 2001; Fig. 32).

Early and Middle Ordovician high-pressure blueschist metamorphism is recorded in Cambrian to Early Ordovician rocks at Port Macquarie (ca. 469 Ma Fukui, 1991; Watanabe et al., 1993; Fukui et al., 1995; Offler, 1999) and along the Peel-Manning Fault (482–467 Ma; Fukui et al., 1995). The latter is contemporaneous with interpreted arc-related plutonism, such as the ca. 480 Ma Attunga gabbro (Fanning et al., 2002). Offler and Shaw (2006) also suggest Late Ordovician arc magmatism (e.g., ca. 445 Ma for hornblende gabbro of calc-alkaline affinity). Similar ages of ca. 436 Ma for tonalite of the Pola Fogal Suite (Kimbrough et al., 1993) supports suggestions of an Early Silurian island arc (Korsch et al., 1990).

Sedimentation and volcanism (but not intrusive magmatism) in eastern Australia largely ended with the Early Silurian Benambran Orogeny (Fig. 32). In the Lachlan, the Benambran Orogeny (e.g., VandenBerg et al., 2000; Glen et al., 2007b) appears to have occurred in two main pulses, ca. 440 and 430 Ma (e.g., Glen et al., 2007b). Deformation associated with the Benambran Orogeny commenced in the Western Lachlan, e.g., Stawell and Bendigo zones, ca. 455–440 Ma (VandenBerg et al., 2000; Gray et al., 2003; Gray and Foster, 2004) and ca. 440 Ma (and slightly older) in other parts of the Lachlan (e.g., Collins and Hobbs, 2001; Gray et al., 2003; Glen et al., 2007b). The resulting Benambran Orogeny affected most of the Lachlan, with the exceptions of the Melbourne Zone and all of Tasmania (VandenBerg et al., 2000; Seymour and Calver, 1995, 1998). The deformation was accompanied by large amounts of shortening, crustal thickening, uplift, and regional metamorphism (e.g., Gray, 1997; Fergusson, 2003). It also marked the end of recorded volcanism in the Macquarie Arc (Crawford et al., 2007a). Significant higher grade metamorphism (e.g., Gray, 1997; Gray et al., 2003) as well as syn-tectonic S-type magmatism (e.g., Collins and Hobbs, 2001), accompanied the orogeny; both are concentrated in two (non-parallel) belts, in the Central (the Wagga-Omeo metamorphic belt), and Eastern Lachlan (shown merged in Fig. 32).

Many of the north Queensland rocks were deformed in the Early Silurian by a shortening event coupled with metamorphism – called the Benambran Orogeny by Fergusson et al. (2007a). Evidence for this deformation is found in the Georgetown region, dated at ca. 430 Ma, synchronous with I-type magmatism (Withnall et al., 1997a; Fergusson et al., 2007a), although it may, in part, be younger (ca. 400 Ma; Withnall et al., 1997a). A similar deformation is recorded in the Charters Towers region, possibly at ca. 440 Ma (Fergusson et al., 2007b). Contractional deformation, related to the Benambran Orogeny(?), is present within the Hodgkinson and Broken River Provinces, but appears to be earlier – probably Late Ordovician. Fergusson et al. (2007a) suggested that island-arc terranes within the Camel Creek Subprovince (Broken River Province) were accreted at this time, which they correlated with the Benambran Orogeny (Figs 31, 33). Garrad and Bultitude (1999) have suggested there may be a time break between Ordovician and Silurian rocks in the Hodgkinson Province that corresponds to uplift (Benambran Orogeny) in the Georgetown region to the west.

The Early Silurian contractional Benambran Orogeny is recorded in the Warburton Basin, Koonenberry Belt and possibly the eastern Thomson Orogen (Gatehouse, 1986; Withnall et al., 1995; Gilmore et al., 2007). Based on metamorphic ages, it is also presumed that the orogeny was experienced by rocks of the central Thomson basement (e.g., Draper, 2006), although the areal extent of the deformation cannot be resolved. The Benambran Orogeny in the New England Orogen apparently coincided with the hiatus between Middle Ordovician and Early Devonian strata at the base of the Tamworth Group (Cawood and Leitch, 1985).

The Benambran Orogeny is interpreted to have resulted in a complex re-arrangement of terranes, particularly in eastern New South Wales and Victoria, but probably also in north Queensland, and perhaps in the southern and eastern Thomson Orogen (Fig. 33). A number of accretion events are inferred to have occurred during the orogeny, though alternate models also exist (e.g., Wyborn, 1992). Interpreted accretion includes the Macquarie Arc terrane and elements of the Adaminaby Superterrane (Glen, 2005; Glen et al., 2007b) and the Benambra Terrane (Willman et al., 2002), as well as the Narooma Terrane (Glen, 2005). Similarly, in north Queensland, it has been suggested that calcalkaline rocks in the Broken River Province represent island arc remnants accreted during the Benambran Orogeny (Fergusson et al., 2007a).

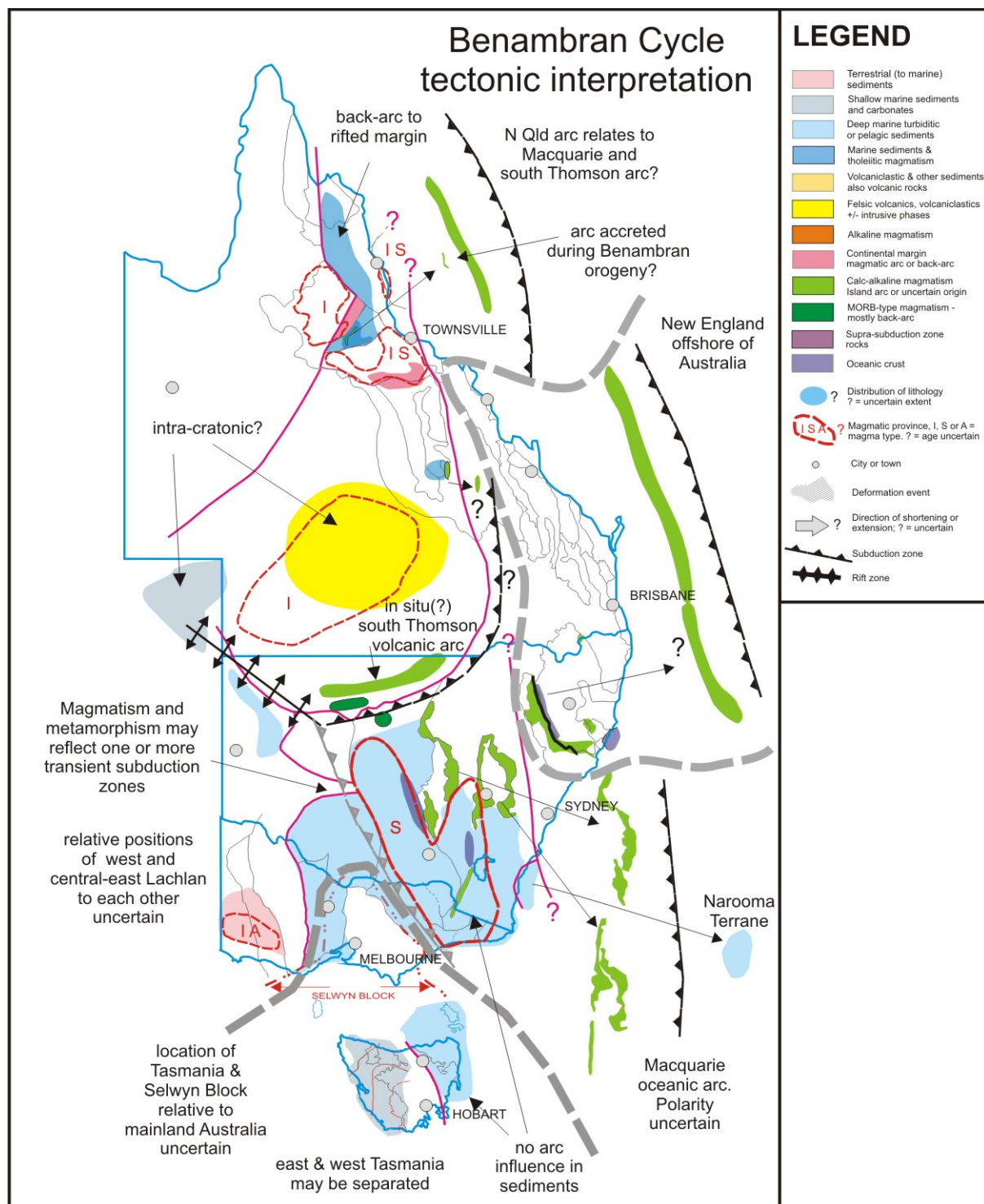


Figure 33. Interpreted tectonic model for the Benambran cycle (ca. 490 Ma – ca. 430 Ma) of eastern Australia. Refer to text for detailed discussion. Location of the Melbourne Zone and Tasmania in this time period is uncertain, as are the relative positions of the West, Central and Eastern Lachlan.

In addition, the interpreted backarc volcanic rocks in the Charters Towers region and Greenvale Subprovince have quite different present-day orientations – ~east-west versus NNE-SSW, respectively (e.g., Bain and Draper, 1997; Figs 31, 33). This suggests either later relative movement between the regions, due to deformation (e.g., Bell, 1980; Fergusson et al., 2007a) and/or perhaps the volcanism formed independently on different crustal fragments. Possible island arc successions occur in the eastern Thomson Orogen (e.g., Withnall et al., 1995). These rocks include fault-bounded serpentinites (Withnall et al., 1995), consistent with accretion. The east-west orientation of possible arc rocks in the southern Thomson Orogen (Fig. 33) also require subsequent accretion, especially if they represent an oceanic arc, as suggested by Watkins (2007). It is, possible however, that they represent an in-situ arc, and some models (e.g., Gray and Foster, 2004), do invoke an east-west arc at this time (Figs 22, 25).

Overall, eastern Australia during the Benambran cycle appears to record both a relatively simple passive-margin to deep marine environment and an oceanic arc environment. These are often depicted as a back-arc/marginal basin behind an oceanic arc and west-dipping slab (e.g., Glen et al., 1998; Li and Powell, 2001; Cayley et al., 2002; Fergusson, 2003; Gray et al., 2003; Gray and Foster, 2004; Glen, 2005; Figs 22, 25). In detail, however, the situation is more complex and controversy exists over the position of terranes, number and location of subduction zones (if any), and mechanisms of terrane accretion, particularly for the Lachlan Orogen (e.g., Coney, 1992; Wyborn, 1992; Gray, 1997; Soesoo et al., 1997; VandenBerg et al., 2000; Collins and Hobbs, 2001; Cayley et al., 2002; Willman et al., 2002; Cas et al., 2003; Fergusson, 2003; Gray et al., 2003; Spaggiari et al., 2003, 2004; Glen, 2005; Glen et al., 2007b). In particular, many models for the Lachlan Orogen invoke a number of subduction zones (e.g., Gray, 1997; Soesoo et al., 1997; Collins and Hobbs, 2001; Fergusson, 2003; Spaggiari et al., 2003, 2004; Fig. 25), to explain the across-orogen variations in magmatism, metamorphism (especially regions of higher grade), deformation (especially structural vergence changes), and the presence of blueschists, as well as the actual width of the Lachlan. These contrast with the tectonically simpler models of VandenBerg et al. (2000), Cayley et al. (2002) and Willman et al. (2002; Fig. 24), for example, which suggest Benambran sedimentation and subsequent deformation in the Western Lachlan occurred in a marginal basin setting behind the Selwyn Block, or the model of Wyborn (1992) with a non-arc setting. Although the actual mechanisms and detail of these reconstructions are beyond the scope of this work, they do have important implications for mineralisation, given that the Benambran event coincides with major lode Au (e.g., Bendigo) and arc-related Cu-Au (e.g., Cadia) mineralisation (e.g., VandenBerg et al., 2000; Crawford et al., 2007a). We have shown a speculative subduction-based model (Fig. 33) which invokes a subduction zone and arc south of the Thomson Orogen, which may link through to the Larapinta Seaway (as also suggested by Gray and Foster, 2004). This arc may join with those suggested for the eastern Thomson Orogen and north Queensland, though the latter, may relate more to the Macquarie Arc and reflect accretion associated with the Benambran Orogeny.

2.4. Silurian to Middle to early Late Devonian

Post-Benambran to Tabberabberan Orogeny: ca. 440 to 380 Ma

The Tabberabberan Cycle (Post-Benambran to Tabberabberan Orogeny) is marked by widespread extensional episodes with accompanying basin formation and widely distributed extrusive and intrusive magmatism. The extension is probably related to significant arc rollback after the Benambran Orogeny (e.g., Glen et al., 2004; Spaggiari et al., 2004). Two orogenies – the Bindian and Tabberabberan - break the Tabberabberan Cycle extension into two episodes. Both are recorded in most regions, with the only significant exceptions being the Melbourne Zone and Tasmania (i.e., the Selwyn Block of Cayley et al., 2002). Significant accretion and amalgamation is also inferred to have occurred during the Tabberabberan Orogeny. This is most evident in the Lachlan Orogen, where, for example, differing geological histories are observed between the Western Lachlan (Whitelaw Terrane) and the Central and Eastern Lachlan (Benambra Terrane), suggesting these regions (terrane of VandenBerg et al., 2000) were separate for most of the cycle (e.g., Gray and Foster, 1997, 2004; VandenBerg et al., 2000; Willman et al., 2002; Spaggiari et

al., 2004). Accretion of island arc terranes to the Gondwana margin in the Middle-Late Devonian within the New England Orogen also occurred during this orogeny.

Within the Central and Eastern Lachlan, post-Benambran extension developed within and across the Ordovician turbidite successions and the Macquarie Arc remnants in New South Wales (Meakin and Morgan, 1999; Lyons et al., 2000; Glen et al., 2007b), and within Ordovician turbidite successions in Victoria (e.g., VandenBerg et al., 2000). This resulted in widespread, largely deep to shallow marine sedimentation, including carbonates, in various basins (Pogson and Watkins, 1998; VandenBerg et al., 2000; VandenBerg, 2003; Colquhoun et al., 2005; Glen, 2005; Lyons et al., 2000; Meakin and Morgan, 1999; Fig. 34). Accompanying magmatism included S- and I-type granites and is often bimodal (Lyons et al., 2000; Collins and Hobbs, 2001; Chappell and White, 1992; Rossiter, 2003; Black et al., 2005; Gray et al., 2003). These basins were inverted during the poorly-defined, latest Silurian (-Earliest Devonian) Bindian Orogeny (ca. 420-410 Ma; e.g., Gray et al., 2003). Renewed extension and development of rift basins continued into the Early Devonian in the Central and Eastern Lachlan following the Bindian Orogeny (e.g., Willman et al., 2002). This resulted in deep to shallow marine sedimentation (including carbonates) and widespread bimodal or felsic volcanism in new and existing basins in the Central and Eastern Lachlan in both Victoria and New South Wales (Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; Willman et al., 2002; Colquhoun et al., 2005; Glen, 2005).

Sedimentation in the Western Lachlan was apparently confined to the Melbourne Zone of Victoria (VandenBerg et al., 2000; Fig. 34), where largely deep marine sedimentation is recorded. Similar extensive deep marine sedimentation occurred in northeastern Tasmania (Seymour and Calver, 1995). The Bindian Orogeny appears to be absent from northeastern Tasmania and the Melbourne Zone of Victoria (Seymour and Calver, 1995; VandenBerg et al., 2000). As a result, sedimentation in both regions is largely continuous throughout this cycle, continuing from Benambran times (e.g., VandenBerg et al., 2000; Fergusson et al., 2003; Seymour and Calver, 1995). In the Melbourne Zone, sedimentation shallows upward to terrestrial sedimentation at the top, and contains evidence for a change in sediment transport direction with the appearance of lithic and volcanoclastic detritus derived from the east (VandenBerg et al., 2000; VandenBerg, 2003). Willman et al. (2002) and VandenBerg (2003) suggested the new source provenance reflects the arrival of the Benambra Terrane, that is, the Benambra and Whitelaw terranes were separate prior to this – a conclusion agreed upon by many workers, regardless of tectonic model (e.g., Gray, 1997; Gray and Foster, 1997, 2004; Fergusson, 2003; Spaggiari et al., 2004).

Widespread felsic-dominated magmatism occurs across the Western, Central and Eastern Lachlan within the Tabberabberan cycle (Lyons et al., 2000; Collins and Hobbs, 2001; Chappell and White, 1992; Rossiter, 2003; Black et al., 2005; Gray et al., 2003; Fig. 34). Ages largely fall between ca. 430 and 390 Ma but continued into the Kanimblan cycle (e.g., Chappell and White, 1992; Gray et al., 2003; Black et al., 2005). The oldest granites in New South Wales and Victoria are dominantly S-types in the Central and Eastern Lachlan (e.g., Collins and Hobbs, 2001; Willman et al., 2002), including a continuation of dominantly S-type magmatism that commenced during the Benambran Orogeny (e.g., Collins and Hobbs, 2001; Figs 32, 34). Early Silurian magmatism appears to be absent from the Western Lachlan (Whitelaw Terrane) in Victoria (VandenBerg et al., 2000; Willman et al., 2002).

Granite ages appear to show a variety of diachronous trends (Fig. 34). In the Eastern Lachlan of New South Wales and Victoria, ages appear to decrease eastwards, from ca. 380 to 360 Ma (Lewis et al., 1994; VandenBerg et al., 2000), possibly reflecting arc rollback. In Victoria, granites mostly decrease in age towards the Melbourne Zone (VandenBerg et al., 2000; Gray et al., 2003; Rossiter, 2003). Granites east and west of the Central Victorian Magmatic Province are largely ca. 420 to 380 Ma in age (VandenBerg et al., 2000; Gray et al., 2003). The youngest rocks occur within the post-Tabberabberan Middle Devonian to Early Carboniferous (ca. 385-350 Ma) Central Victorian Magmatic Province (VandenBerg et al., 2000; Rossiter, 2003). A similar diachronous trend is evident in Tasmania, where granite ages record a pronounced westward younging from ca. 400-375 Ma, pre-, syn-, and post-tectonic granites in the northeast to post-tectonic granites, ca. 370-350 Ma, in western Tasmania (Black et al., 2005; Fig. 34).

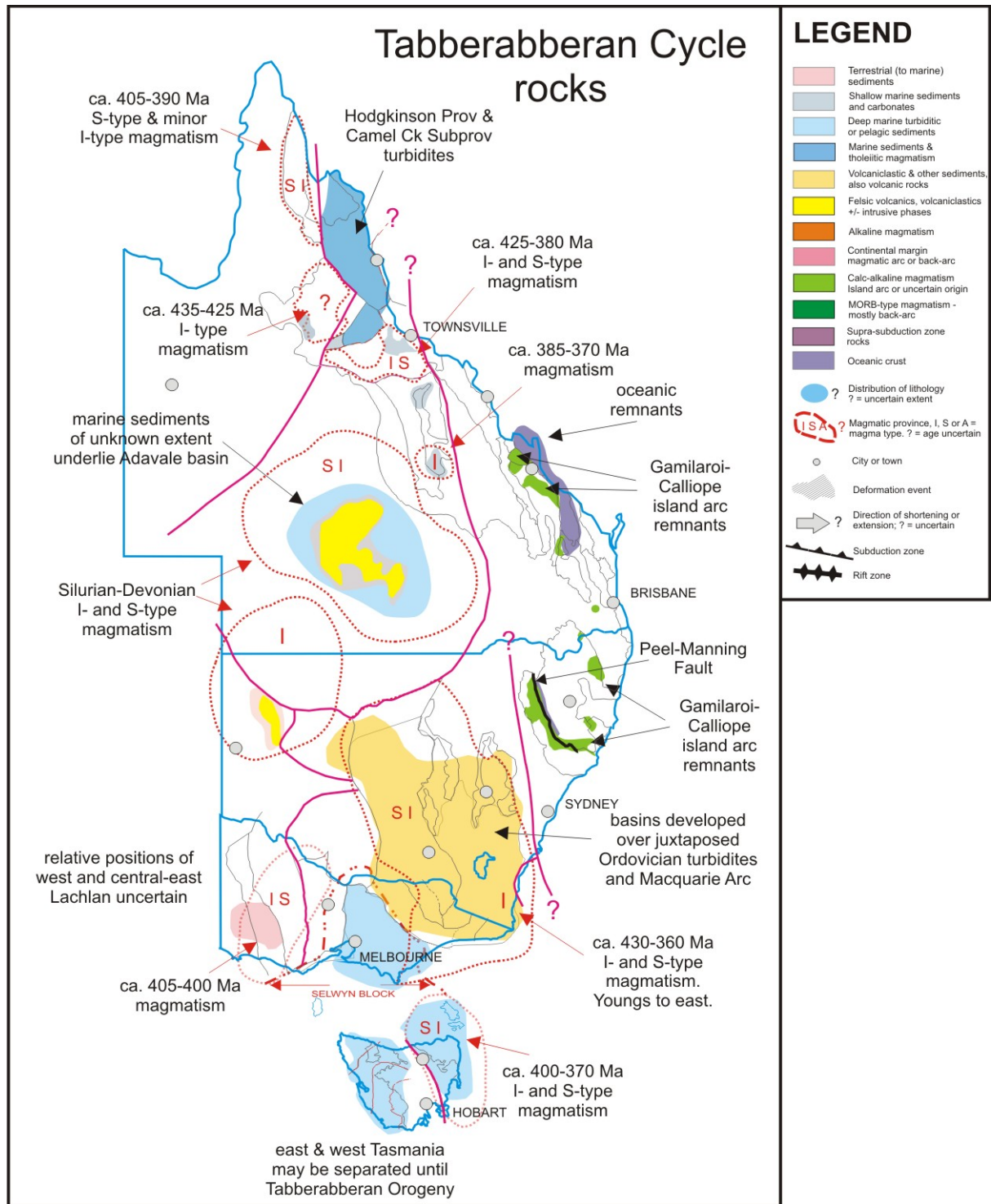


Figure 34. General distribution of Silurian to early Late Devonian (ca. 440 to 380 Ma), Tabberabberan cycle rocks in eastern Australia. Refer to text for discussion.

Tabberabberan Cycle rocks in the southern Delamerian Orogen include terrestrial to marine sedimentation in western Victoria, largely in the Grampians-Stavely Zone (VandenBerg et al., 2000), and deep-water, clastic-dominated sedimentation in Western Tasmania (Seymour and Calver, 1995, 1998; Fig. 34). Sediments in Western Victoria were apparently deformed at ca. 420–410 Ma (Bindian Orogeny?) and are overlain by post-deformation Early Devonian volcanic rocks (VandenBerg et al., 2000) with associated plutonism. The Bindian Orogeny appears to be absent in western Tasmania, although hiatuses which may correspond to this orogeny are recorded (Seymour and Calver, 1995, 1998).

The Tabberabberan Cycle in north Queensland is characterised by extensive, pre- and post-Bindian, sedimentation in the Hodgkinson and Broken River Provinces (along the eastern and southeastern margins of the Proterozoic Georgetown block), post-Bindian sedimentation in the Charters Towers region, and minor pre-Bindian sedimentation in the Georgetown region (Fig. 34). Sedimentation in these regions includes marine siliciclastic sediments and carbonates, and, locally abundant, pre-Bindian tholeiitic mafic volcanism (Arnold and Fawcner, 1980). Sediment provenance is dominantly cratonic (e.g., Bultitude et al., 1997) but does include volcanoclastic material, some of which is older (for example, Garrad and Bultitude (1999) record dacitic clast ages of 465 Ma in the Hodgkinson Province). The geodynamic setting for this sedimentation is controversial. Models include both backarc or forearc deposition, as well as a rifted continental margin (see summaries in Arnold and Fawcner (1980) and Garrad and Bultitude (1999)). Arnold (in Arnold and Fawcner, 1980), Henderson et al. (1980) and Henderson (1987), amongst others, suggested that the Hodgkinson and Broken River Province sedimentation was part of a forearc basin and accretionary wedge. This model, however, has difficulties explaining the tholeiitic volcanism, especially in the Hodgkinson Province. The latter is more consistent with rift or backarc models (e.g., Fawcner in Arnold and Fawcner, 1980; Bultitude et al., 1997), though the magmatism may simply reflect accreted oceanic crust fragments in the accretionary wedge model. Resolution of the geodynamic setting is critical to understanding the tectonics of the widespread Pama Province magmatism. The latter forms an extensive quasi-continuous belt around the Hodgkinson and Broken River Provinces, from Charters Towers in the south, north to Cape York. Pama Province magmatism in the region is diachronous. It ranges from ca. 430–420 Ma (syn- and post-Benambran) in the Georgetown region, to ca. 425 to 405 Ma (and younger) in the Charters Towers region, to ca. 410 to 395 Ma in the Coen region. As pointed out by Champion and Bultitude (2003), these age differences are also matched by changes in geochemical signature. Notably, the early Pama magmatism (in the Georgetown region) does have geochemical signatures more consistent with arc magmatism.

Sedimentation and associated extrusive and intrusive magmatism in the Thomson Orogen and Koonenberry region also occurred within two episodes, also probably in response to extension following the Benambran and Bindian orogenies (e.g., Olgers, 1972; Neef and Bottrill, 1991; Murray, 1994, 1997a; McKillop et al., 2005; Withnall et al., 1995; Gilmore et al., 2007). Post-Benambran, Late Silurian to Early Devonian marine and terrestrial sediments, of cratonic and/or volcanic provenance, are also recorded in the Koonenberry region (Neef and Bottrill, 1991), and in the Thomson Orogen (Olgers, 1972; Murray, 1994, 1997a; Withnall et al., 1995; Fig. 34). These are associated with mafic and felsic volcanic rocks in the Koonenberry region (Neef and Bottrill, 1991) and the Anakie Inlier (Withnall et al., 1995), and felsic intrusives in the Thomson Orogen basement (e.g., Murray, 1994) and Koonenberry region (Gilmore et al., 2007; Fig. 34). Both Murray (1994) and Thalhhammer et al. (1998) have suggested continental settings. Widespread felsic intrusive magmatic rocks of this age (ca. 425–405 Ma), belonging to the Pama Province (Bain and Draper, 1997) occur within the Charters Towers region (Hutton et al., 1997; Fig. 34). Renewed extension, following the Bindian Orogeny, produced the Early to Late Devonian, terrestrial to shallow marine Adavale Basin in the central Thomson Orogen (McKillop et al., 2005), terrestrial to shallow marine sedimentation in the Burdekin Basin (Charters Towers region; Hutton et al., 1997), and Early to Middle Devonian quartz-rich sedimentation in the Koonenberry region (Neef, 2004). The Adavale Basin is considered to have formed in a continental setting, possibly as an intracontinental volcanic rift (Murray, 1994; McKillop et al., 2005). Felsic intrusive magmatism accompanied extension in the Thomson Orogen (e.g., Murray, 1994; Evans et al., 1990; Hutton et al., 1997). McKillop et al. (2005) suggest that extension may have been the result of far field events, such as subduction further to the east (in the New England Orogen).

The Tabberabberan cycle in the New England Orogen is characterised by a convergent margin phase, represented by the Late Silurian to Middle Devonian Gamilaroi-Calliope island(?) arc and an enigmatic collisional phase (Tabberabberan Orogeny), associated with the possible accretion of these island arc terranes in the Middle-Late Devonian. Elements of a Late Silurian-Middle Devonian arc occur in fault-bounded blocks along the western flank of the Peel-Manning Fault in the southern New England Orogen, and in the vicinity of the Yarrol Belt in the northern part of the orogen (Fig. 34). The arc successions consisting of volcanoclastic, extrusive and intrusive rocks of the Gamilaroi Terrane and Calliope Arc, are interpreted to have formed as a single intraoceanic island arc (low-K calc-alkaline signature) (van Noord, 1999; Offler and Gamble, 2002; Murray and Blake, 2005), though Morand (1993a) suggested a primitive continental arc environment was possible. Volcanic rocks in the Gamilaroi Terrane lie below the Middle Devonian-Carboniferous strata of the Tamworth Belt. The Calliope Arc succession, host to the Mount Morgan gold-copper deposit (~380 Ma), is overlain by strata of the Yarrol Belt and is intruded by the Mount Morgan Trondhjemite (381±5 Ma; Golding et al., 1994), which has an arc or rifted-arc geochemistry (Murray, 2003). Cawood and Flood (1989) and Offler and Gamble (2002) have suggested that the arc developed above a west-dipping subduction zone, although Aitchison and Flood (1995) argued for east-dipping subduction. The Silverwood Group and Willowie Creek beds, have been interpreted as a southern continuation of the Calliope Arc (Day et al., 1978). Island arc magmatism is also recorded along the Peel-Manning Fault (Early Silurian ca. 425 Ma hornblende cumulates and diorites; Sano et al., 2004) and in the Marlborough Block (Middle Devonian). In the latter, calcalkaline basalts, dolerites and gabbros and fault-bounded blocks in the Marlborough ophiolite (380±19 Ma) have trace element data suggestive of an intra-oceanic island arc (Bruce and Niu, 2000). The arc was probably built on a Neoproterozoic oceanic crust, lay only a short distance offshore, and was accreted to Gondwana (Bruce and Niu, 2000; Murray, 2007). Late Silurian to Devonian ocean basin sedimentation is also recorded in the New England Orogen, in the Woolomin Terrane (428-380 Ma chert and basalt; Aitchison et al., 1992), and in the Coastal Block (e.g., ~418 Ma-~360 Ma Doonside Formation; Fergusson et al., 1993; Fig. 34).

The Bindian Orogeny (ca. 420-400 Ma), although largely poorly defined and variably developed, appears to be present within the Lachlan, Delamerian, Thomson and north Queensland orogenies, but not the New England Orogen (Fig 35). The orogeny is most evident in the Central and Eastern Lachlan in Victoria and New South Wales, where it has been suggested to be transpressive and to have resulted in significant strike-slip movement between the western and central Lachlan (VandenBerg et al., 2000; Willman et al., 2002; Glen, 2005), with possibly up to 600 km of dextral movement (e.g., Willman et al., 2002; Fig. 24). Other authors have questioned this, and it is possible that deformation of this age may relate more to continuing subduction-accretion effects if the multiple subduction zone models of Gray (1997), Gray and Foster (1997), Soesoo et al. (1997), Spaggiari et al. (2003, 2004; Fig. 25), are correct. This does not necessarily negate strike-slip effects in the Central and Eastern Lachlan. Bindian deformation in far eastern Victoria appears to relate more to east-west contraction (e.g., Willman et al., 2002). Bindian-aged deformation is also present in parts of the Delamerian Orogen, for example, the Tabberabberan cycle sediments in Western Victoria were deformed at ca. 420-410 Ma (VandenBerg et al., 2000; Fig. 35). The Bindian Orogeny appears to be absent, however, in western Tasmania, although hiatuses which may correspond to this orogeny are recorded (Seymour and Calver, 1995, 1998). Local folding and metamorphism, probably with associated felsic magmatism, took place in the Koonenberry region and the Thomson Orogen during the Bindian Orogeny (e.g., Thalhhammer et al., 1998; Fig. 35). The orogeny there may have been diachronous. Deformation in the Koonenberry region is recorded as Late Silurian to Early Devonian (Gilmore et al., 2007), while it appears to be late Early Devonian in the Thomson Orogen, constrained by ca. 408 and 402 Ma volcanic rocks in the post-Bindian Adavale Basin (McKillop et al., 2005). Bindian-aged deformation - ca. 410-400 Ma - is also recorded in the Georgetown, Coen and Charters Towers regions of north Queensland, where it coincides with part of the extensive Pama Province magmatism in those regions (Fig. 35). Bultitude et al. (1997) record a change in sedimentation in the Hodgkinson Province in the Late Lochkovian (ca. 412 Ma), and Withnall et al. (1997b) record a hiatus in sedimentation at this time in the Graveyard Creek Subprovince. Both suggested these changes were related to hinterland uplift (due to Bindian orogeny?). Sedimentation also appears to recommence at this time in the Charters Towers region.

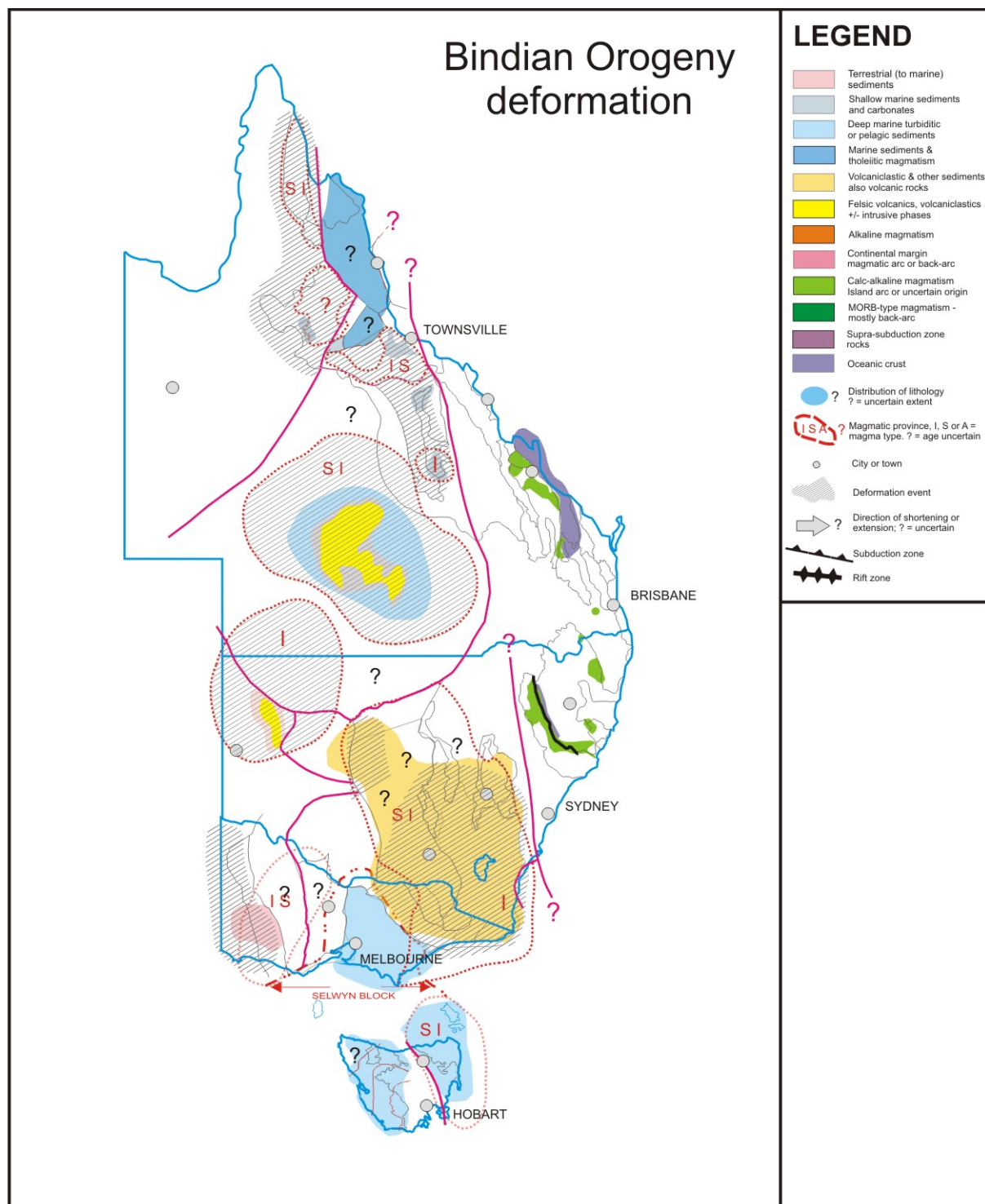


Figure 35. The generalised (approximate) distribution of deformation (cross-hatching) during the Bindian Orogeny deformation (ca. 420 to 400 Ma) as defined by outcrop and drill hole information – actual extent of deformation may be greater and more continuous. The intensity of deformation is variable within the areas indicated. See text for data sources. Refer to [Figure 34](#) for additional geological explanations.

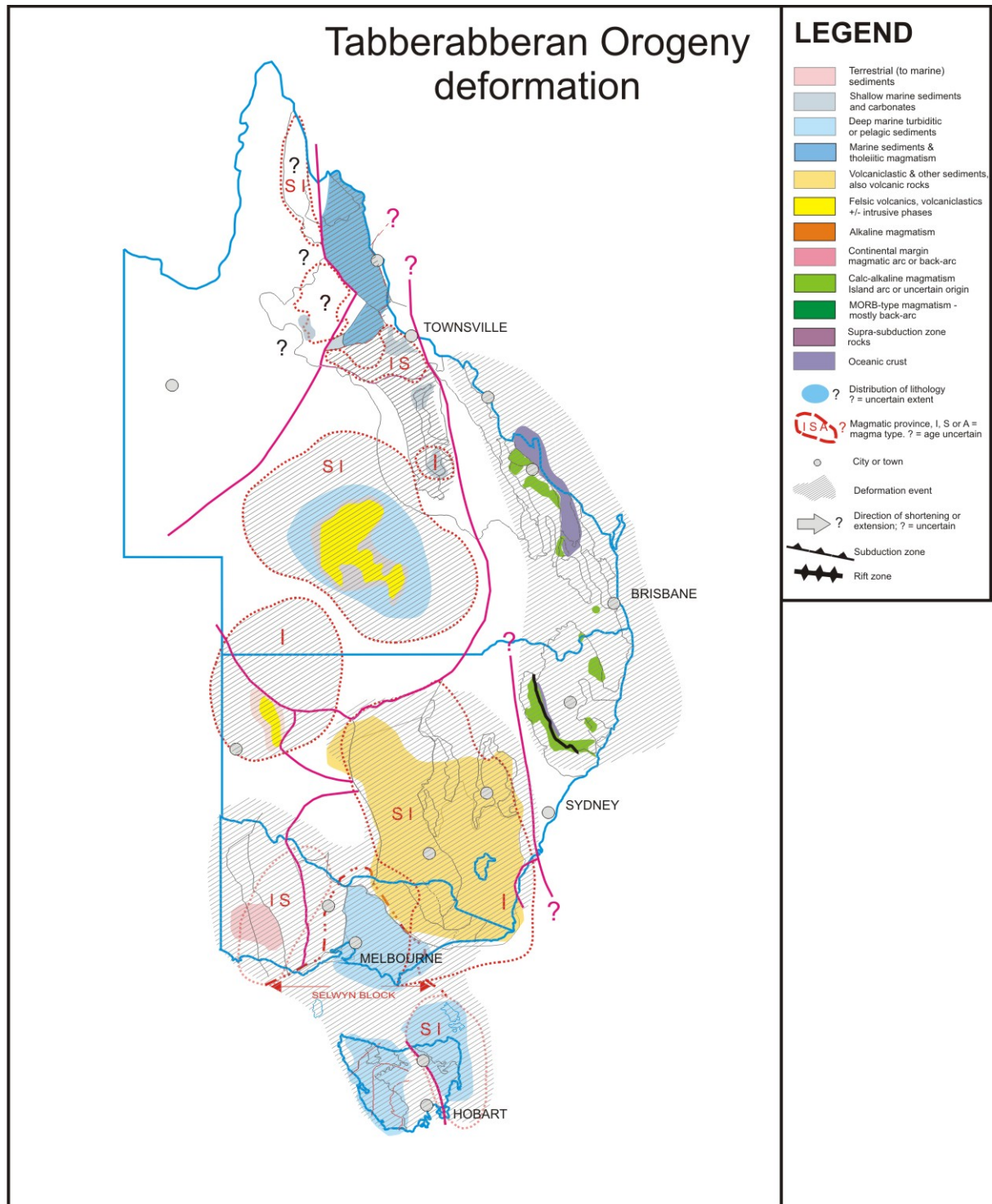


Figure 36. The generalised (approximate) distribution of deformation (cross-hatching) during the Tabberabberan Orogeny (ca. 390 to 380 Ma) as defined by outcrop and drill hole information – actual extent of deformation may be greater and more continuous. The intensity of deformation is variable within the areas indicated. See text for data sources. Refer to [Figure 34](#) for additional geological explanations.

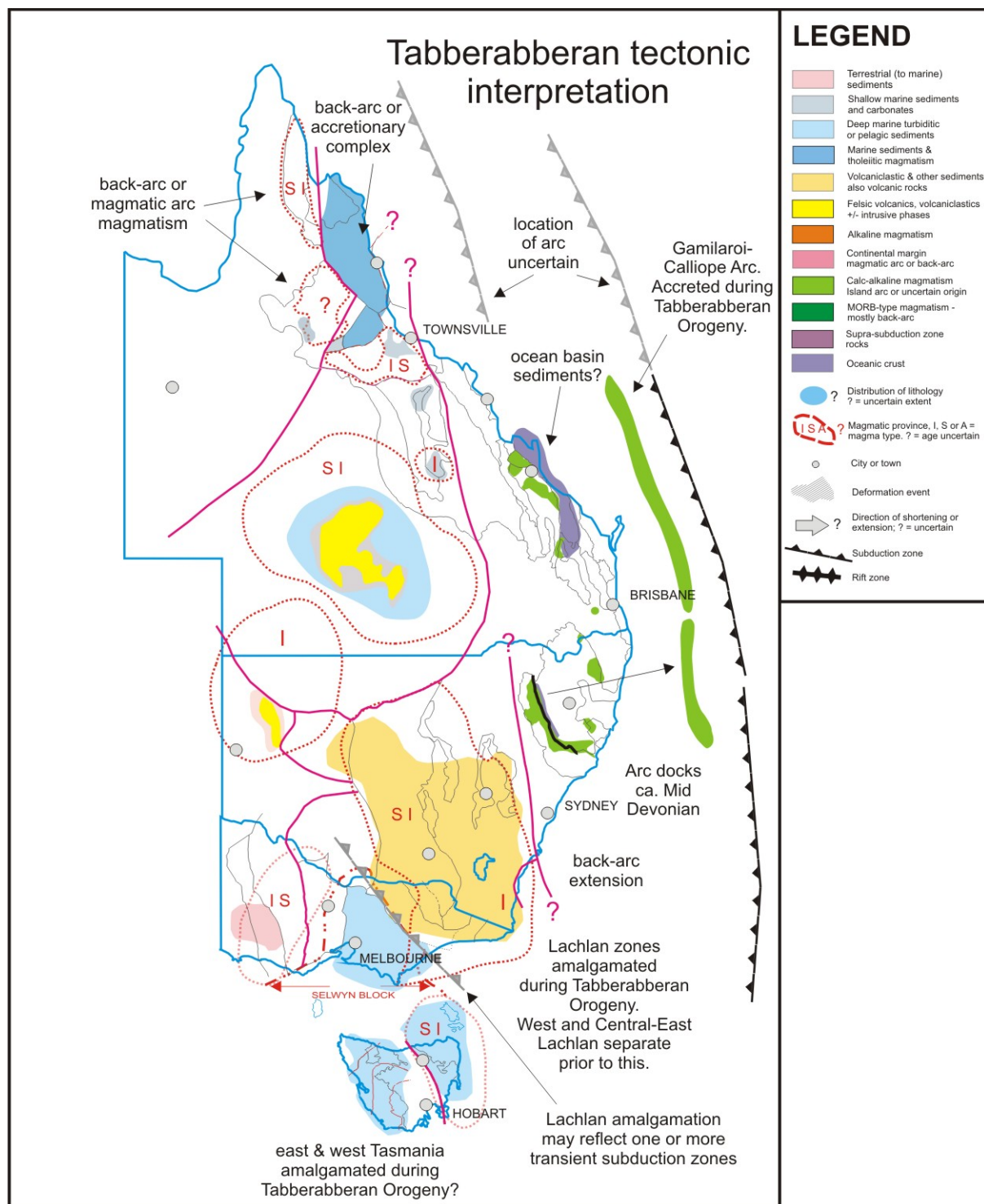


Figure 37. Interpreted tectonic model for the Tabberabberan cycle (ca. 440 to 380 Ma) of eastern Australia. Refer to text for detailed discussion.

The Tabberabberan Orogeny affected most of eastern Australia (Fig. 36). Within the Lachlan Orogen, the ca. 390-380 Ma dominantly east-west contractional Tabberabberan Orogeny (e.g., Gray and Foster, 1997, 2004; Spaggiari et al., 2003, 2004), effectively cratonised the whole orogen, and has been interpreted to have been responsible for amalgamation of the terranes and zones of the Lachlan (Gray and Foster, 1997, Soesoo et al., 1997; VandenBerg et al., 2000; Willman et al., 2002; Spaggiari et al., 2003, 2004; Seymour and Calver, 1995; Black et al., 2005; Figs 24, 25). Although dominantly east-west, a component of north-south contraction is also recorded as part of this orogeny within the Lachlan Orogen (e.g., VandenBerg et al., 2000), perhaps consistent with oblique terrane amalgamation (e.g., Spaggiari et al., 2003). Deformations around this age in north Queensland are best represented in the Broken River and Hodgkinson provinces and also in the Charters Towers region, where they include time breaks and slight angular unconformities (Withnall and Lang, 1993; Bultitude et al., 1997; Fig. 36). Henderson (1987) suggested that this event resulted in the cessation of deep-marine sedimentation in the Camel Creek Subprovince, and produced the Middle Devonian angular unconformity observed in the Graveyard Creek Subprovince. Garrad and Bultitude (1999) recorded east to north-east thrusting and north-northwest-trending shear zones in the Hodgkinson Province, which they suggested were of Late Devonian age, possibly related to basin inversion. Given the strong commonalities between the Broken River and Hodgkinson Provinces it is possible deep-water marine sedimentation ceased simultaneously in both. The Tabberabberan Orogeny in the Koonenberry region resulted in east-northeast – west-southwest contractional deformation at ca. 395 Ma (Mills and David, 2004; Neef, 2004; Fig. 36). Deformation of this age in the Thomson Orogen (often called the Alice Springs Orogeny (2)) is either poorly developed or difficult to distinguish from other events (Withnall et al., 1995; Hutton et al., 1997). In the Adavale Basin, the orogeny appears to have produced an unconformity between terrestrial and overlying shallow marine sedimentary rocks, and a possible change to restricted basin conditions in the late Middle Devonian (McKillop et al., 2005).

The Tabberabberan Orogeny is also recorded in the New England Orogen, although there is no general agreement as to its effects and timing. In the southern New England Orogen, Flood and Aitchison (1992) suggested that the Calliope-Gamilaroi intraoceanic arc accreted to the Australian plate in the Late Devonian (Figs 36, 37), based on the first appearance of distinctive, westerly-derived quartzite clasts in the uppermost Devonian Keepit Conglomerate of the Tamworth Group, considered to have been derived from the Lachlan Orogen. This is controversial, however, and according to Glen (2005), no clear evidence of contractional deformation related to arc accretion has been identified. In the northern NEO, supporters of the intraoceanic 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). In contrast, others (e.g., Leitch et al., 1992; Morand, 1993b; Bryan et al., 2003) have argued that the unconformity is low angle, deformation was minor and that there was no break at the base of the overlying Yarrol Belt.

Like the Benambran cycle (ca. 490 to 430 Ma), it would appear that the Tabberabberan cycle (ca. 430 to 380 Ma) records a relatively simple overall backarc environment behind a subduction zone to the east within the New England Orogen (e.g., Gray, 1997; Glen et al., 1998; Cayley et al., 2002; Gray and Foster, 2004; Glen, 2005; Collins and Richards, 2008; Fig. 37). Like the Benambran cycle, it is clear that there is unresolved tectonic complexity within the Tabberabberan cycle. Much of this concerns the Lachlan Orogen, and reflects the continuation of the varied tectonic regimes invoked for the Benambran cycle. These have been discussed in this and the previous sections (see Figs 24, 25; see also Gray et al., 2003); most attention here is paid to the tectonic drivers of the Tabberabberan Orogeny, and the origins of the widespread intrusive magmatism.

The large areal distribution of magmatism in eastern Australia during the Tabberabberan cycle (Fig. 34) is problematical and has led many authors to speculate on tectonic scenarios, especially for the Lachlan Orogen. Suggested tectonic environments for the latter include multiple subduction zones (e.g., Collins and Hobbs, 2001; Soesoo et al., 1997; Gray and Foster, 1997), delamination (Collins and Vernon, 1994), mantle plumes/upwellings (e.g., Wyborn, 1992; Cas et al., 2003), as well as backarc extension and episodic contraction related to variable arc rollback (VandenBerg, 2003). The multiple subduction models, especially the divergent double subduction model of Soesoo et al. (1997), have the advantage of potentially

explaining the distribution of, and the observed diachronous trends shown by, the Lachlan granites, especially in the western and central Lachlan. The dominantly felsic nature of much of the magmatism (e.g., Chappell & White, 1992), however, is not consistent with multiple subduction zones (though cf. Collins and Hobbs, 2001). Similarly, plume models can explain the widespread nature of the magmatism but not the diachronous trends of the granites. It is noted that magmatism of this age in the Thomson Orogen and Koonenberry region (Fig. 34), although poorly understood and with little geochronological control, also appears to occupy a similar east-west extent as the Lachlan Orogen, suggesting that backarc to behind-arc processes (not multiple subduction zones) may perhaps be sufficient to explain the width of magmatism in the Lachlan (Fig. 37). Differing tectonic models also exist for the widespread Pama Province magmatism in north Queensland and Charters Towers (Fig. 34). Resolving the tectonic interpretation for the Pama Province magmatism depends significantly on geodynamic models for the Hodgkinson and Broken River Provinces, which have been interpreted as either forearc or backarc (Fig. 37). In forearc models, the magmatic belt is interpreted as the magmatic arc (e.g., Henderson, 1987), consistent with some (e.g., Champion and Bultitude, 2003), but not all (e.g., Blewett et al., 1997), of the granite geochemistry. Similarly, the presence of tholeiitic magmatism within the Hodgkinson Province (Arnold and Fawcner, 1980; Henderson, 1987; Bultitude et al., 1997), is perhaps more consistent with the backarc model but not inexplicable with a fore arc model. The diachronous nature of the Pama Province magmatism and the corresponding changes in geochemical signature (e.g., Champion and Bultitude, 2003), may perhaps be best interpreted as a switch from an early forearc environment (ca. 430 Ma) to a backarc environment (ca. 420-400 Ma and younger?), similar to the model of Collins and Richards (2008) for the Lachlan Orogen.

Interpretations for the drivers of the Tabberabberan contraction are varied. Most work has concentrated on the Lachlan Orogen and conclusions largely reflect the different tectonic models inferred for the region. Gray and co-workers (e.g., Gray, 1997; Gray and Foster, 1997, 2004; Soesoo et al., 1997; Spaggiari et al., 2003, 2004) suggested a collisional event that was (at least partly) related to the closure of a marginal basin (effectively the Melbourne Zone) and an end to double divergent subduction (e.g., Gray and Foster, 1997; Soesoo et al., 1997; Figs 23, 25, 37). Conversely, Willman et al. (2002) and Cayley et al. (2002) suggested the Tabberabberan Orogeny was responsible for ending the relative strike-slip movement of the Whitelaw and Benambra terranes, and reflected docking of the two terranes (Fig. 24). As discussed in the previous section, each model better explains certain, but not all, aspects of the Tabberabberan cycle geology. Importantly, most models indicate that the Western and Central-Eastern Lachlan were separate for much of the Tabberabberan cycle and that they were amalgamated during the Tabberabberan Orogeny (e.g., Gray, 1997; Gray and Foster, 1997, 2004; VandenBerg et al., 2000; Willman et al., 2002; Fergusson, 2003; Spaggiari et al., 2004). A similar argument seems valid for Tasmania. Detrital zircon data from sediments of this age in northeastern Tasmania indicate no apparent sourcing of material from western Tasmania, which led Black et al. (2004) to suggest that northeastern and western Tasmania were also separate at this time, such that in Tasmania the Tabberabberan deformation (ca. 388 Ma; Black et al., 2005) may relate to docking of northeastern Tasmania to western Tasmania (Black et al., 2004). It would appear, therefore, that at least part of the Tabberabberan Orogeny relates to amalgamation of the Lachlan Orogen. It is also probable, however, that this deformation may also relate to docking of the Calliope-Gamilaroi arc in the New England Orogen at this time, although as discussed earlier, the effects and timing of any amalgamation are not clear cut or agreed (e.g., Flood and Aitchison, 1992; Leitch et al., 1992; Morand, 1993; Bryan et al., 2003; Murray et al., 2003; Glen, 2005). What does appear evident, however, is that the end result of the Tabberabberan Orogeny was a crustal configuration for eastern Australia not far removed from that we see today.

2.5. Late Devonian to Early Carboniferous:

Post-Tabberabberan to Kanimblan Orogeny: ca. 380 Ma - ca. 350 Ma

The interpreted accretion of the Gamilaroi-Calliope island arc to the Gondwana margin in the Tabberabberan Orogeny resulted in the initiation of the New England Orogen as an Andean-style

continental margin, with a westerly dipping subduction zone (Cawood, 1982; Murray et al., 1987), that continued throughout the Kanimblan cycle (ca. 380 to 350 Ma). During this period, the now largely cratonic Lachlan, Delamerian, Thomson and North Queensland Orogens were marked by widespread extension, rifting and accompanying basin formation, and significant extrusive and intrusive magmatism (e.g., VandenBerg et al., 2000; Lyons et al., 2000; Glen, 2005). Like earlier orogenic cycles, this extension is thought to reflect behind-arc processes, including a backarc basin in the north, related to renewed rollback in the subduction zone and arc within the New England Orogen (e.g., Glen, 2005; Collins and Richards, 2008). This extension ended with the Kanimblan Orogeny (ca. 350 Ma), although the arc itself – the Connors-Auburn Arc – continued throughout the Kanimblan cycle and into the next cycle, from the Late Middle Devonian to the Late Carboniferous (ca. 380 Ma - ca. 305 Ma; e.g., Roberts et al., 1995; Murray, 2003). The subsequent history of the New England Orogen, therefore, records an event history in many ways distinct from the remainder of the Tasmanides (e.g., Glen, 2005).

The Andean-style continental margin in the New England Orogen resulted in the development of a magmatic arc in the west (Connors-Auburn Arc), flanked by a forearc basin and an accretionary wedge in the east (Cawood, 1982; Murray et al., 1987; Fig. 38). In the northern NEO, the continental magmatic arc is represented by granitic and mafic to silicic rocks of the Connors-Auburn Arc (Murray, 1986). Although magmatism probably began earlier, the preserved record shows that igneous activity in both arches commenced ca. 350 Ma – late in the Kanimblan Cycle, with the main pulse of granite formation between ca. 324-313 Ma in the Auburn Arch and ca. 316-305 Ma in the Connors Arch (Murray, 2003). In the southern NEO, the magmatic arc is now either buried or has been removed. Its presence is inferred from large Late Devonian olistromal blocks of andesitic volcanic rocks in the inboard part of the forearc basin (Brown, 1987) and by Late Devonian to Late Carboniferous sandstones largely of arc-derived volcanoclastic composition (Cawood, 1983; Korsch, 1984). The Tamworth Belt in the southern NEO and the Yarrol Belt in the northern NEO, as well as intervening blocks, are usually interpreted to be parts of a continuous forearc basin (Cawood and Leitch, 1985; Korsch et al., 1990; Murray et al., 2003; Fig. 38). These Late Devonian to Late Carboniferous successions consist predominantly of continental to shallow marine clastic sediments, including limestone, with a provenance predominantly from the volcanic arc to the west (Cawood, 1983; Korsch, 1984; Yarrol Project Team, 1997). The late Devonian rocks of the Yarrol Province are considered to represent the transitional change from an intraoceanic setting to a continental margin setting in the Carboniferous (Murray and Blake, 2005). Volcanism and deposition continued until the Late Carboniferous (Roberts et al., 2004). The Woolomin, Central and Coffs Harbour blocks in the southern NEO and the Coastal, Yarraman, North D'Aguilar, South D'Aguilar and Beenleigh blocks in the northern NEO are interpreted as a once continuous accretionary wedge (Fig. 38) that grew oceanwards by accreting trench-fill volcanoclastic turbidites (derived from a magmatic arc) and minor amounts of oceanic crust (e.g., Korsch et al., 1990). A backarc basin occurs in the north (Drummond Basin), with the Late Devonian of the Lachlan Orogen being the backarc equivalent in the south.

The earliest Kanimblan cycle rocks in the Lachlan Orogen consist of Middle to Late Devonian sediments and A-type and bimodal extrusive and intrusive magmatism (e.g., Meakin and Morgan, 1999; Lyons et al., 2000; Lewis et al., 1994; Wormald et al., 2004), probably related to initiation of post-Tabberabberan extension and associated rifting. This was followed by Lachlan-wide (New South Wales and Victoria) late Middle to Early Carboniferous, clastic, mostly continental, sedimentation, including red beds, of the 'Lambie facies', reflecting continuing, more widespread extension (Lewis et al., 1994; Warren et al., 1995; Meakin and Morgan, 1999; Lyons et al., 2000; VandenBerg et al., 2000; Cas et al., 2003; Glen, 2004; Fig. 38). Middle Devonian to earliest Carboniferous intrusive I-, S- and A-type magmatism, ca. 380 to 350 Ma (Chappell and White, 1992; Gray et al., 2003; Wormald et al., 2004; Black et al., 2005), occurred throughout this cycle.

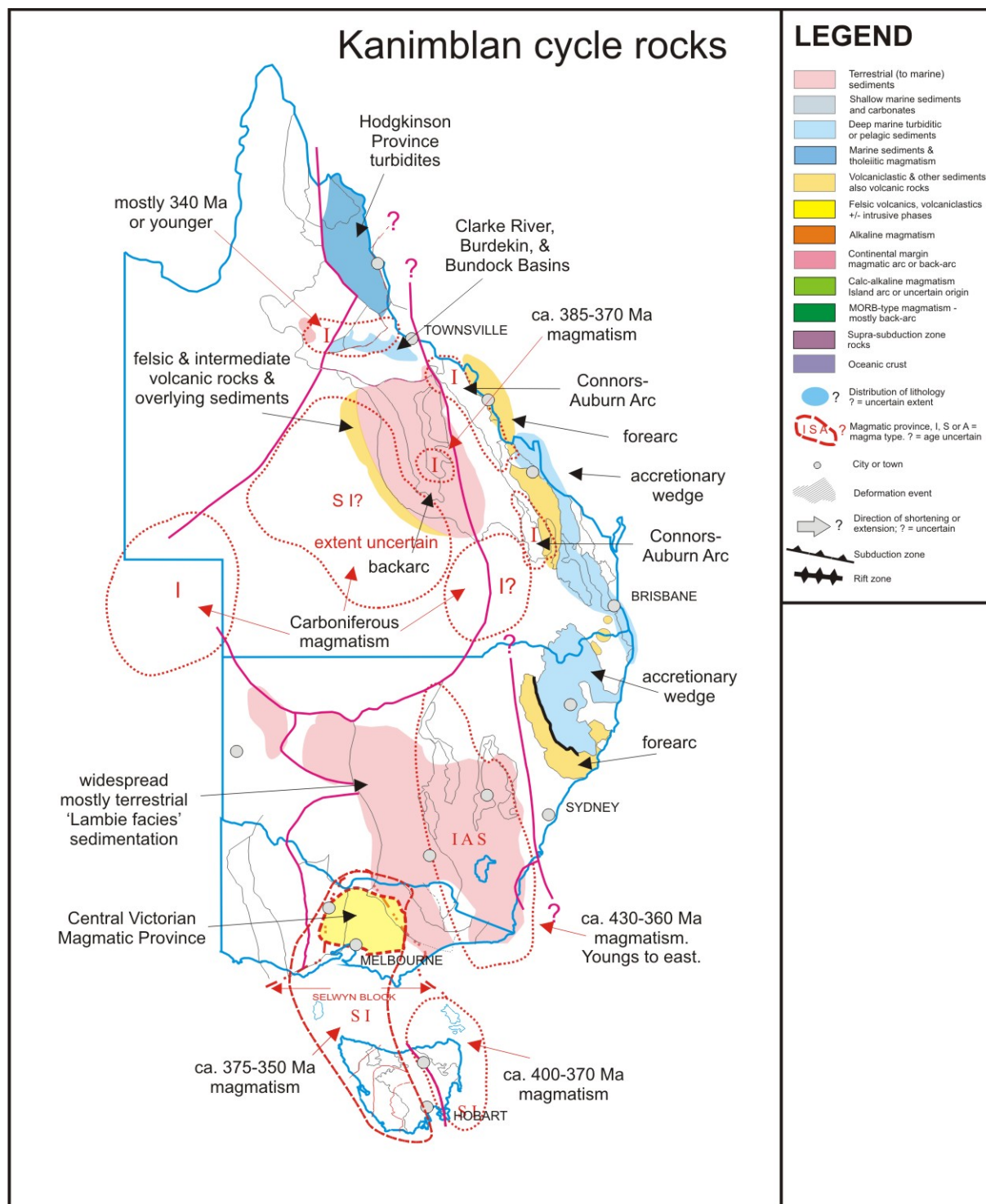


Figure 38. General distribution of Late Devonian to Early Carboniferous Kanimblan cycle (c. 380 Ma to c. 350 Ma) rocks in eastern Australia. Refer to text for discussion. The New England Orogen continued in this configuration (arc, forearc and accretionary wedge) until the Late Carboniferous.

As noted above, the granites in the Lachlan Orogen appear to show a variety of diachronous trends (Fig. 38). In the Eastern Lachlan of New South Wales and Victoria, ages appear to decrease eastwards, to ca. 380 to 360 Ma (Lewis et al., 1994; VandenBerg et al., 2000; Glen, 2005), most probably reflecting subduction zone rollback. In Victoria, granites appear to mostly decrease in age towards the Melbourne Zone, from both east and west sides (VandenBerg et al., 2000; Gray et al., 2003; Rossiter, 2003). The youngest rocks occur within the post-Tabberabberan Middle Devonian to Early Carboniferous (ca. 385-350 Ma) Central Victorian Magmatic Province (VandenBerg et al., 2000; Rossiter, 2003). A similar diachronous trend is evident in Tasmania, where granite ages record a pronounced westward younging in granite age from ca. 400-375 Ma, pre-, syn-, and post-tectonic granites in the northeast to post-tectonic granites, ca. 370-350 Ma in western Tasmania (Black et al., 2005; Figs 34, 38). There is little evidence for arc-related magmatism during this period in the Lachlan, with most evidence clearly indicating that the arc was located further east in the New England Orogen (e.g., Meakin and Morgan, 1999; Glen, 2005; Collins and Richard, 2008).

The Kanimblan cycle in north Queensland consists of largely non-volcanic (cratonic provenance), terrestrial and lesser marine sedimentation across all regions, best preserved in the lower successions of the Bundock, Clarke River and Burdekin basins of the Broken River and Charters Towers regions. Minor andesitic volcanism is recorded in the Georgetown region (Withnall et al., 1997a), and minor volcanoclastic input is recorded in several of the regions.

During this cycle, both terrestrial and marine sedimentation, often with accompanying volcanism, occurred in the Thomson Orogen and in the Koonenberry region, largely in response to intracratonic extension following the Tabberabberan Orogeny, but also in response to backarc extension behind the arc in the New England Orogen (e.g., Neef and Bottrill, 1991; Murray, 1994; Withnall et al., 1995; Henderson et al., 1998; Draper et al., 2004; McKillop et al., 2005; Gilmore et al., 2007). Backarc extension resulted in rifting and initiation of the Drummond Basin in the latest Devonian (Henderson et al., 1998). The Drummond Basin contains a thick succession of continental and lesser marine sediments and volcanic rocks (Olgers, 1972; Hutton et al., 1998; Henderson et al., 1998), with the lowermost units being latest Devonian to Early Carboniferous syn-rift related volcanics rocks and associated marine to terrestrial volcanoclastic sediments (Olgers, 1972; Henderson et al., 1998). In the Thomson Orogen, felsic and lesser intermediate and mafic magmatism accompanied extension episodically throughout this cycle. This includes intrusive and related extrusive magmatism in the Anakie Inlier, prior to Drummond Basin formation, regionally extensive, Late Devonian to Early Carboniferous, felsic magmatism in the Thomson Orogen, including within the Drummond Basin (e.g., Olgers, 1972; Murray, 1994), and Early Carboniferous granites in the Thomson Orogen basement and Warburton Basin strata (e.g., Murray, 1994). The Late Devonian-Early Carboniferous backarc silicic magmatism in the Drummond Basin at this time may be related to episodes of silicic magmatism in the New England Orogen (e.g., Bryan et al., 2004).

The extension was ended by the Early Carboniferous (ca. 360-340 Ma) contractional east-west Kanimblan Orogeny (e.g., Gray, 1997; Meakin and Morgan, 1999; VandenBerg et al., 2000; Gray et al., 2003; Glen, 2005). This orogeny folded and inverted Kanimblan cycle and older rocks. It occurred across the Lachlan Orogen, into the Delamerian Orogen (e.g., Gilmore et al., 2007), but is best expressed in the Eastern Lachlan (Gray, 1997; Gray et al., 2003; Glen, 2005). As outlined by Willman et al. (2002), the post-Tabberabberan sedimentation and volcanic rocks in the Lachlan Orogen overlie major faults and interpreted suture zones belonging to the Tabberabberan Orogeny, with little evidence for significant later reactivation. Tabberabberan deformation in north Queensland was minor in nearly all areas, with the exception of the Hodgkinson Province, where significant east-west shortening and further basin inversion occurred (Garrad and Bultitude, 1997). This deformation immediately pre-dates the commencement of the voluminous and widespread extrusive and intrusive magmatism of the Kennedy Province.

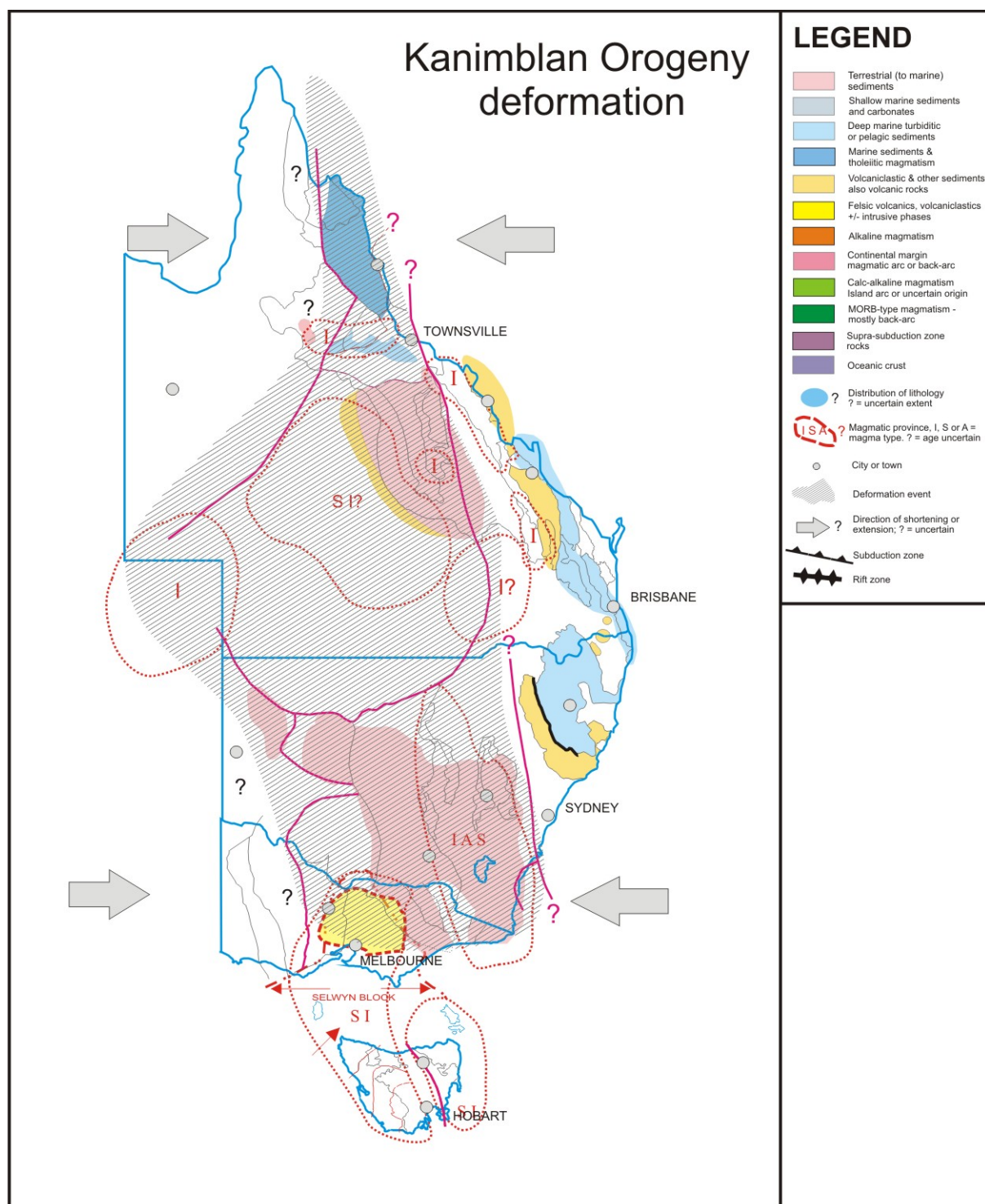


Figure 39. Generalised (approximate) distribution of deformation (cross-hatching) within the Kanimblan Orogeny (ca. 360 to 350 Ma) as defined by outcrop and drill hole information – actual extent of deformation may be greater and more continuous. The intensity of deformation is variable within the areas indicated. See text for data sources. Refer to [Figure 38](#) for additional geological explanations.

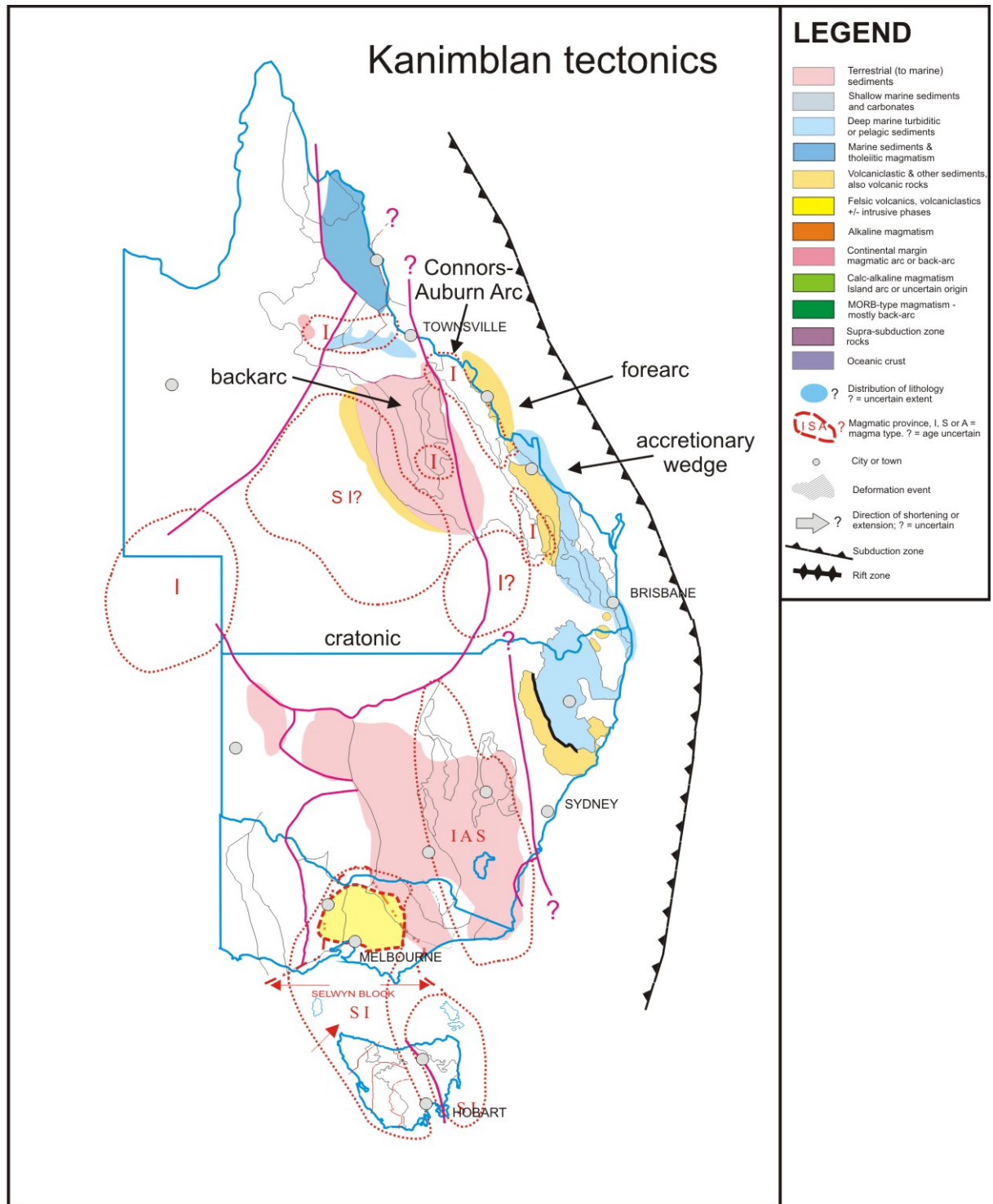


Figure 40. Interpreted tectonic model for the Kanimblan cycle (ca. 380 to 350 Ma) of eastern Australia. Refer to text for detailed discussion.

The Early to ?Middle Carboniferous Kanimblan Orogeny, or Alice Springs Orogeny (3) as it is known in the Thomson Orogen, (also referred to as the Quilpie Orogeny in the Devonian basins of Queensland; Finlayson et al., 1990; Finlayson, 1990), produced a major episode of faulting and deformation in the Koonenberry Belt, slight contraction in the Drummond Basin and regional-scale folding and subsequent erosion in the Adavale Basin (e.g., Olgers, 1972; Neef, 2004; Gilmore et al., 2007). Deformation in the Drummond Basin is recorded by an unconformity between the Drummond and overlying strata in the Galilee Basin (Scott et al., in prep). This deformation event is suspected to have driven regional-scale, southward thrusting of the Thomson over the Lachlan Orogen (Korsch et al., 2007).

2.6. Middle Carboniferous to late Early Permian

Post-Kanimblan to orocline formation - ca. 350 Ma to ca 270 Ma

This cycle is largely defined based on the tectonic evolution of the New England Orogen. It includes the continuation of Connors-Auburn Arc in the New England Orogen until its apparent end in the Late Carboniferous (ca. 305 Ma; Cawood, 1982; Murray et al., 1987), and, the ensuing extensional, backarc phase from the Late Carboniferous to late Early Permian (ca. 305 Ma to ca. 270 Ma; Holcombe et al., 1997a, b). This interval is characterised by extension, accompanied by widespread magmatism and sedimentation, including initiation of the Sydney-Gunnedah-Bowen Basin system (Fig. 41). Extension was terminated by deformation and formation of the Texas and Coffs Harbour Oroclines (Fig. 42).

The Andean-style continental margin in the New England Orogen appears to have been minimally affected by the Kanimblan Orogeny and continued into the post-Kanimblan with little change to the magmatic arc (Connors-Auburn Arc), forearc basin or accretionary wedge (Cawood, 1982; Murray et al., 1987; Figs 38, 41). The orogeny did, however, result in changes to the backarc, terminating sedimentation in the Drummond Basin (Olgers, 1972), for example. The main recorded pulses of granite formation in the Connors-Auburn Arc occurred within this cycle - between ca. 324-313 Ma in the Auburn Arch and ca. 316-305 Ma in the Connors Arch (Murray, 2003). As for the Kanimblan cycle, although there is no record of the magmatic arc in the southern New England Orogen (either buried or removed), its presence is inferred from voluminous arc-derived volcanoclastics in the forearc (Cawood, 1983; Korsch, 1984). Volcanism and deposition, predominantly of continental to shallow marine clastic sediments, continued within the forearc basin (e.g., Tamworth Belt, Yarrol Belt; Cawood, 1983; Korsch, 1984; Cawood and Leitch, 1985; Korsch et al., 1990; Yarrol Project Team, 1997; Murray et al., 2003; Figs 38, 41) until the Late Carboniferous (Roberts et al., 2004). Similarly, oceanward accretion of trench-fill volcanoclastic turbidites (derived from a magmatic arc) and minor amounts of oceanic crust (basalt, chert, mudstone), continued within the accretionary wedge east of the arc (e.g., Korsch et al., 1990; Figs 38, 41).

The New England Orogen underwent a transition from active accretion in the mid-Carboniferous to widespread extension through the Late Carboniferous-Early Permian (Leitch, 1988; Holcombe et al., 1997a), interpreted to reflect the eastward retreat of the subducting slab, and migration of the volcanic arc offshore (Holcombe et al., 1997a; Roberts et al., 2004). Mechanisms invoked for arc migration include slab breakoff (Caprarelli and Leitch, 2001) and rollback (Jenkins et al., 2002). By the Early Permian, much of the New England Orogen was in an extensional continental backarc setting (Holcombe et al., 1997a, b; Fig. 41). Deposition of bimodal volcanics, volcanoclastics and siliciclastic sedimentary rocks occurred in numerous extensional basins (Leitch, 1988; Roberts et al., 1996; Korsch et al., 1998, in press a, b; Fig. 43), overlying older forearc and accretionary wedge successions. Extension is also recorded by the emplacement of granites into the former accretionary wedge (Glen, 2005).

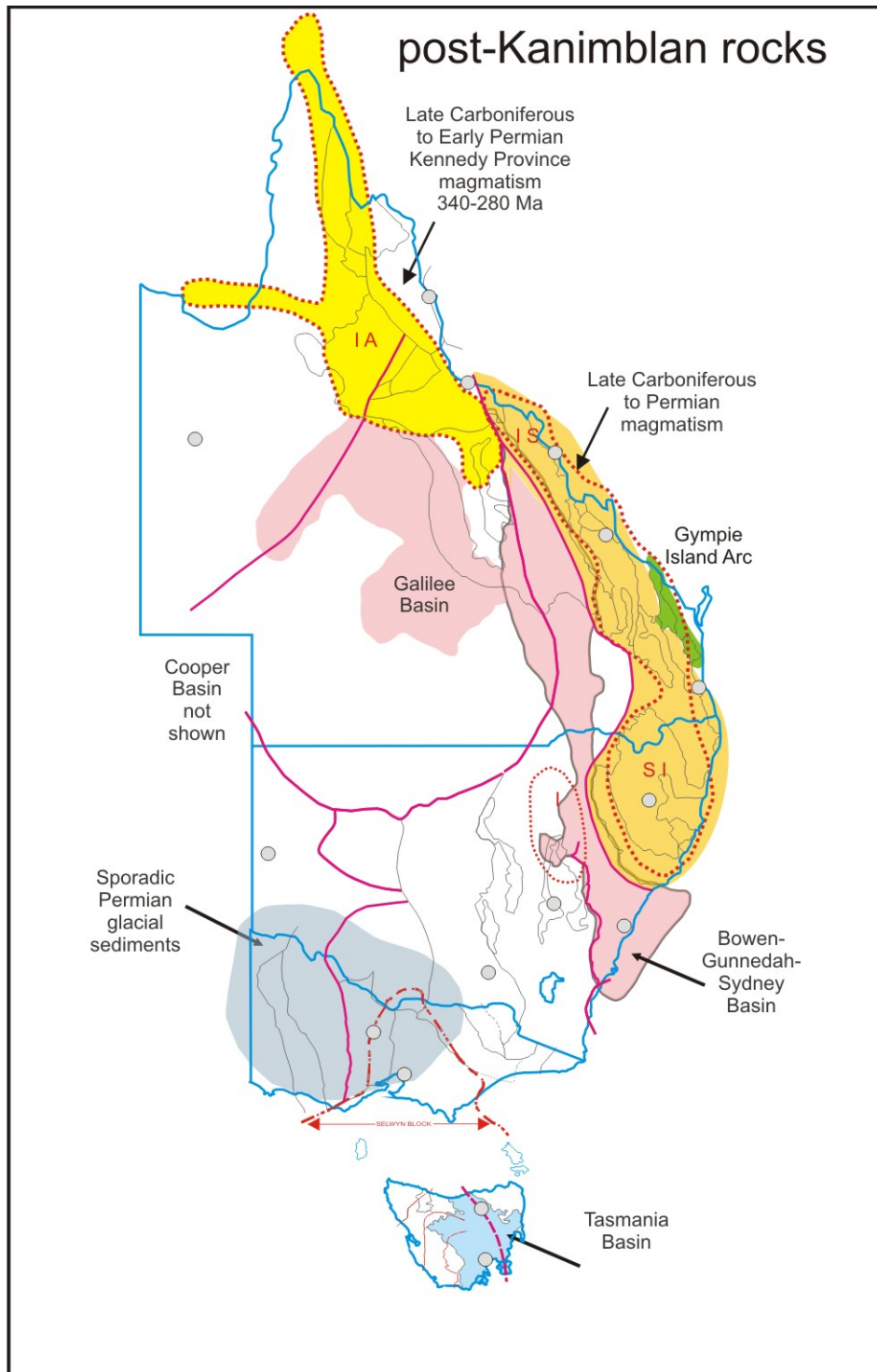


Figure 41. General distribution of Early Carboniferous to mid Permian post-Kanimblan cycle rocks in eastern Australia. Refer to text for discussion. Note: The configuration shown for the New England Orogen reflects the latter (Late-Carboniferous and younger) part of the cycle. Prior to this, the New England Orogen consisted of a continuation of the magmatic arc, forearc and accretionary wedge, as for the Kanimblan cycle (see [Figure 38](#)).

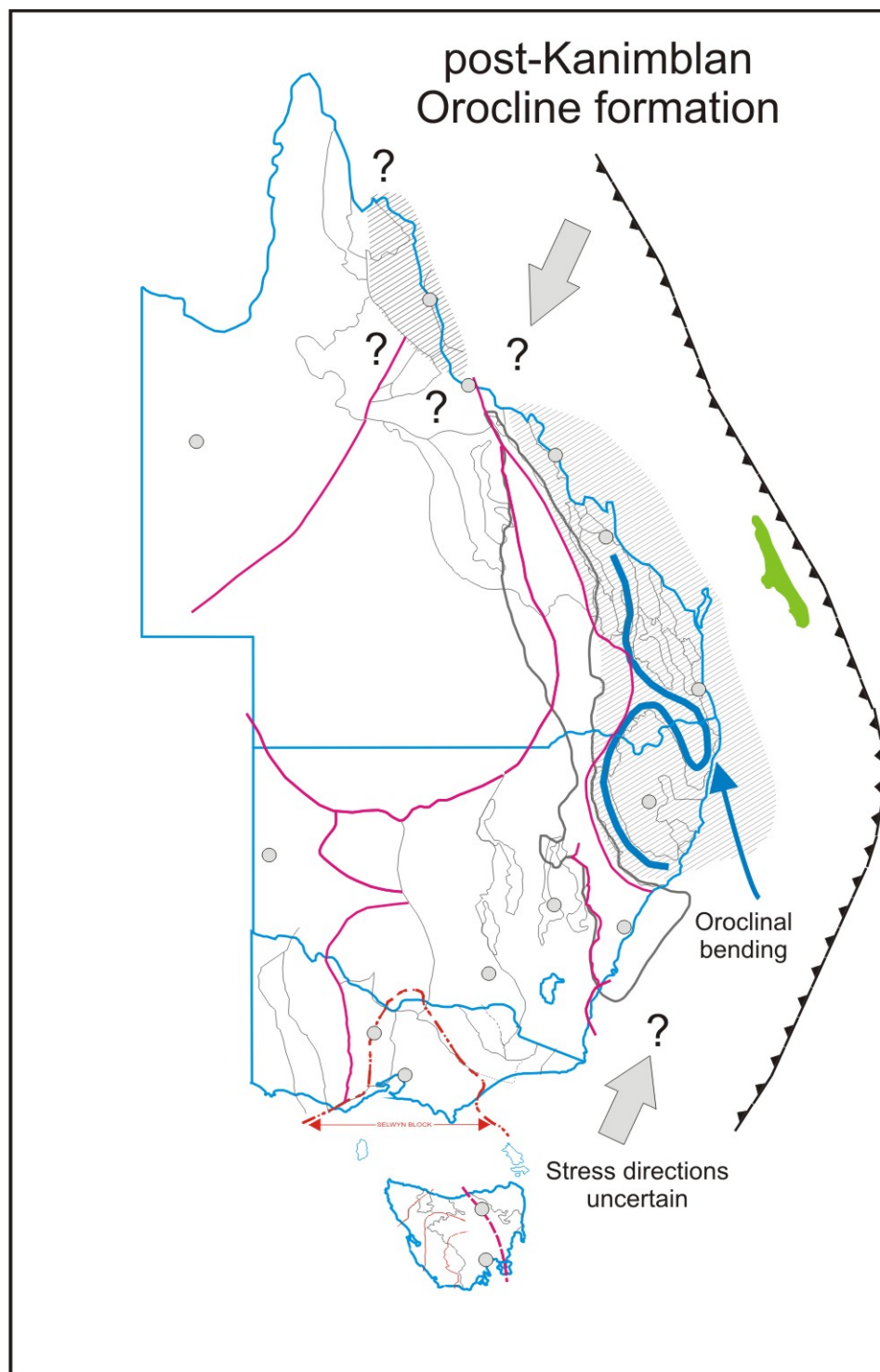


Figure 42. Generalised (approximate) distribution of deformation within the post-Kanimblan orocline forming event in the New England Orogen – actual extent of deformation may be greater and more continuous. The intensity of deformation is variable within the areas indicated. See text for data sources.

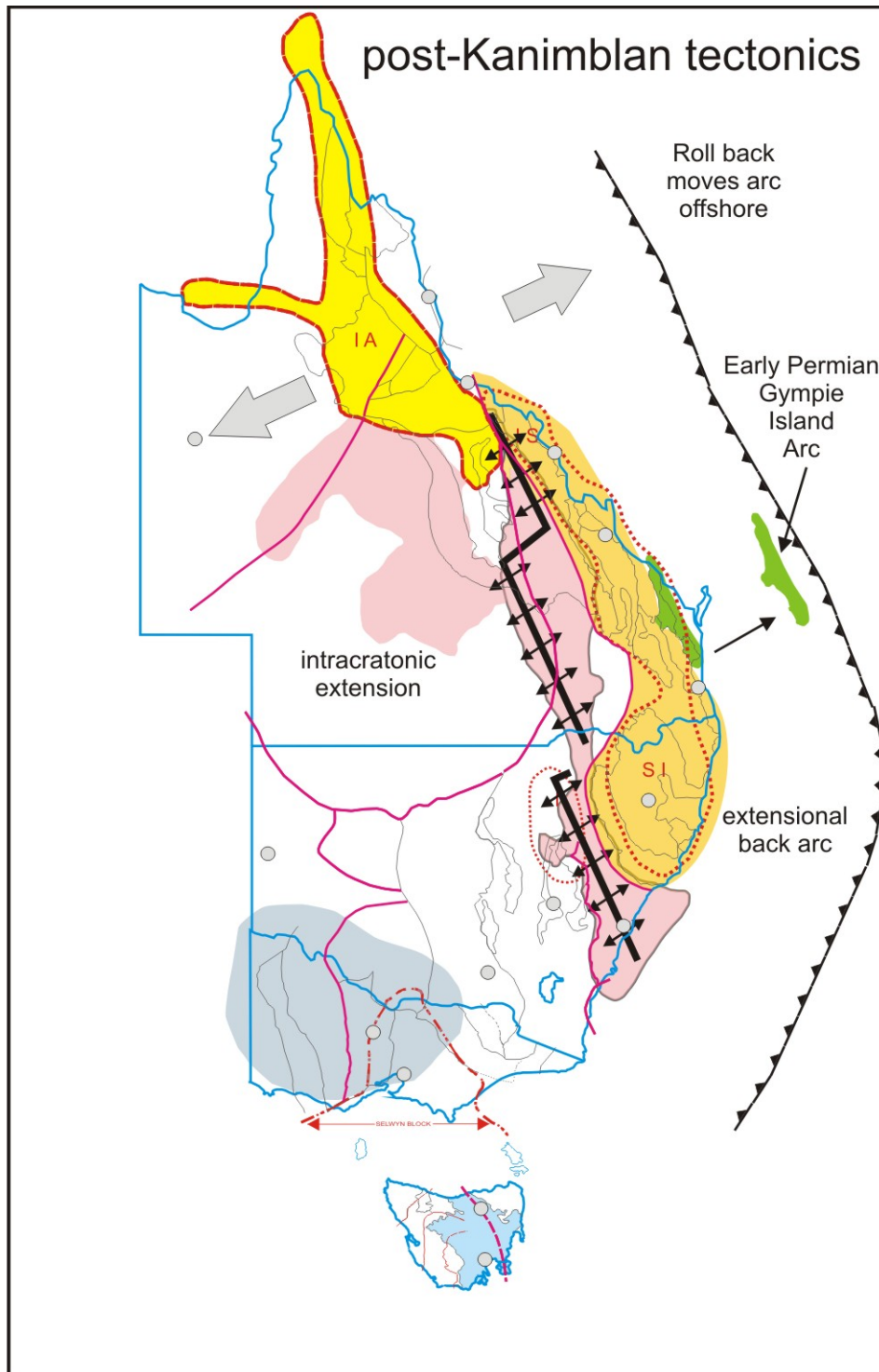


Figure 43. Interpreted tectonic model for the post-Kanimblan cycle of eastern Australia. Refer to text for detailed discussion.

Cessation of subduction (~305 Ma; synchronous with cessation of deposition in the Tamworth Belt; Roberts et al., 1995, 1996) and an eastward shift in magmatism at ca. 300 Ma, probably via rollback, resulted in generation of the S-type Hillgrove Suite in an extensional backarc environment in the southern New England Orogen (~302 Ma; Collins et al., 1993; Kent, 1994; Jenkins et al., 2002). S-type plutons in the northern New England Orogen (North D'Aguilar Block) are also possibly related to the earliest phase of extension (e.g., 309 ±4 Ma Gallangowan Granite; 310 ±4 Ma Yabba Creek Granite; Holcombe et al., 1997a).

Early Permian subsidence in the western regions of the backarc gave rise to the Sydney-Gunnedah-Bowen Basin System. The event that initiated the formation of these basins was extensional (at ~305 Ma; Denison Event; Korsch et al., in press a, b), and the continental crust was stretched to form the significant Early Permian East Australian Rift System, as defined by Korsch et al. (1998; in press a). Deposition within the Sydney-Gunnedah-Bowen Basin System was largely controlled by events within the orogen to the east, particularly the switch (later in the Permian) of the basin from back-arc to a foreland basin system (Roberts et al., 1996; Korsch et al., 1998; see below).

An inferred Early Permian backarc setting for the New England Orogen and Sydney-Gunnedah-Bowen Basin requires that an arc environment existed outboard at this time (Figs 41, 43). The Gympie Terrane contains Early Permian submarine and subaerial volcanic rocks which have been suggested to have a chemical signature indicative of a juvenile island-arc terrain, isolated from the influence of continental crust (Sivell and Waterhouse, 1988; Sivell and McCulloch, 1997, 2001; Figs 41, 43). These rocks are intimately associated with primitive oceanic backarc basalts (Sivell and McCulloch, 1997) suggesting that a well-developed backarc separated the primitive Gympie arc from the New England Orogen, that is, the intraoceanic Gympie island arc was located east of the continental margin of Gondwana in the Early Permian (Fig. 43). Detrital zircon data from Gympie (Korsch et al., in press c) suggest that the Gympie arc was attached back to eastern Australia at the end of the Permian or the start of the Triassic.

Also in the Early Permian, but probably after extension, oroclinal bending of the forearc and accretionary wedge successions (~285-265 Ma) produced the Texas and Coffs Harbour Oroclines (Korsch and Harrington, 1987; Murray et al., 1987; Fig. 42). Drivers for the oroclinal folding are varied, including transform faulting, transtension and transpression (e.g., Korsch and Harrington, 1987; Murray et al., 1987; Fergusson and Leitch, 1993; Korsch et al., 1990; Offler and Foster, 2008), although, as summarised by Offler and Foster (2008), most models invoke dextral movement (Fig. 42). The timing of oroclinal bending has also been the subject of much debate (e.g., see Murray et al., 1987; Fergusson and Leitch 1993; Korsch and Harrington, 1987; Offler and Foster, 2008); we place oroclinal development after cessation of Early Permian backarc extension but prior to the main phase of the Hunter-Bowen Orogeny. Recent work by Offler and Foster (2008) suggests orocline deformation commenced at ca. 276 Ma.

The Kanimblan Orogeny was the terminal event in the Lachlan Orogen (e.g., Pogson and Watkins, 1998), and subsequent geology there relates largely to the New England Orogen to the east. From the Middle Carboniferous to Permian, Eastern Australia was dominated by tectonic extension and rifting, and formation of intracratonic basins. Within Victoria and New South Wales, the only significant magmatic event was the late Early Carboniferous to Late Carboniferous (ca. 340-310 Ma) intrusive magmatism occurring as a north-northwest belt in the northeastern Lachlan (e.g., Pogson and Watkins, 1998; Meakin and Morgan, 1999; Fig. 41). This magmatism consists of the I-type, mostly felsic, granites of the Bathurst Batholith and the Gulgong Suite (Bathurst basement terrane of Chappell et al., 1988; Fig. 41). They are of similar age and geochemistry to volcanic and intrusive rocks in the New England Orogen (e.g., Chappell et al., 1988; Pogson and Watkins, 1998), and may relate to continental arc formation, although Meakin and Morgan (1999) suggest emplacement in an extensional environment, presumably behind the continental arc of the New England Orogen. The eastern part of the Lachlan Orogen is overlain by the latest Carboniferous to Triassic Sydney-Gunnedah-Bowen basin system, which also, at least, initially developed as a backarc rift behind the New England Orogen (e.g., Korsch et al., in press a). The basin rocks overlie the Carboniferous granites of the Bathurst-Gulgong area. Of similar age to the Sydney-Gunnedah-Bowen Basin System are sediments of the Parmeener Supergroup in the Tasmania Basin in Tasmania), which developed over the

suture (Tamar Fracture) between western and northeastern Tasmania (Seymour and Calver, 1995, 1998; Fig. 41). These consist of a lower succession of glacial and marine sediments and an upper succession (Late Permian and younger) of nonmarine sediments, including coal measures (Seymour and Calver, 1995, 1998). Remnants of possibly more widespread Permian glacial and marine sedimentation are also recorded (outcrop and sub-surface, e.g., beneath the Murray Basin) in Victoria and southern New South Wales (O'Brien et al., 2003).

This period in north Queensland and in the northern Thomson Orogen (Charters Towers region and northern Drummond Basin) is characterised by the commencement of the widespread and voluminous extrusive and intrusive Kennedy Province magmatism, plus associated, mostly minor sedimentation (Fig. 41). As documented by numerous authors (e.g., Richards et al., 1966) this magmatism is crudely diachronous, commencing earlier in the Georgetown, Broken River and Charters Towers regions (ca. 340–335 Ma), and younging to mid Permian in the Hodgkinson Province. There are also accompanying changes in geochemistry. Magmatism in the mid to Late Carboniferous is almost exclusively I-type, along with some mantle-derived magmatism. In the Early Permian, magmatism switched to A- and I-type in the Georgetown and western Hodgkinson regions, and to S- and I-type in the central and eastern Hodgkinson (Champion and Bultitude, 2003). The tectonic regime for this magmatism is not well understood. Despite the strong crustal input into the magmatism, it is generally thought to be broadly arc-related (e.g., Champion and Bultitude, 2003, though cf. Murray et al., 1987), probably in an extensional backarc or behind-arc position relative to the continental arc in the New England Orogen (Fig. 43). This is consistent with the younging of magmatism to the east matching the inferred migration of the Connors-Auburn Arc eastwards at this time (Holcombe et al., 1997a; Jenkins et al., 2002; Roberts et al., 2004).

Deformation in this cycle in north Queensland appears to have occurred at least twice. There is a widespread but minor north-south contraction, thought to be mid to Late Carboniferous in age, and commonly equated with the Alice Springs Orogeny, found in the Broken River, Hodgkinson and Georgetown regions (Withnall et al., 1997a, b; Bultitude et al., 1997). In addition, there is a more significant deformation in the Early Permian, possibly related to an early phase of the Hunter-Bowen Orogeny. This penetrative east-west shortening deformation is best documented in the Hodgkinson Province (e.g., Davis, 1994; Davis et al., 1996; Garrad and Bultitude, 1999; Fig. 42).

From the Late Carboniferous to Early Permian, tectonic extension and rifting initiated extensive intracratonic basin formation in eastern Australia (e.g., Cooper and Galilee basins; Bain and Draper, 1997; Fig. 43). Continental margin extension in the Early Permian led to formation of the Bowen-Gunnedah-Sydney basins in a backarc setting (Korsch et al., 1998; in press; Figs 41, 43), consistent with the presence of latest Carboniferous to Early Permian bimodal and calcalkaline volcanics at the base of the Bowen Basin succession (e.g., Green et al., 1997b; Withnall et al., in prep). Sedimentation in the Bowen Basin was contiguous with the Gunnedah Basin and the former unconformably overlies the Drummond Basin (Green et al., 1997a; Scott et al., in prep). By the Early Permian, the Bowen Basin, together with the Gunnedah and Sydney Basins, formed the 'East Australian Rift System' (Korsch et al., 1998; Fig. 43). Although extension was located on the continental margin, basins such as the Galilee and Cooper formed on the craton further west during the late Carboniferous to Permian at this time (Draper and McKellar, 2002; Figs 41, 43). These contain widespread terrestrial and glacial sediments, including coal measures (e.g., Scott et al., 1995; Draper, 2002a, b, 2004; Gray and McKellar, 2002). A connection between the Cooper and Galilee basins meant the two basins experienced related sediment deposition (Scott et al., 1995). It has been suggested that deformation by the Alice Springs Orogeny (3) caused convective downwelling and regional downwarp of the Drummond Basin, resulting in the formation of troughs and depressions in the Galilee Basin (Jackson et al., 1981; Middleton and Hunt, 1989).

2.7. Late Early Permian to Middle Triassic:

The Hunter-Bowen Orogeny - ca. 270 Ma to ca. 230 Ma

This cycle in eastern Australia is largely defined by the New England Orogen and associated backarc basins. The cycle commenced in the late Early to early Late Permian (~265-262 Ma) when a magmatic arc was apparently re-established along the Paleo-Pacific continental margin of Australia and the previous backarc switched from an extensional to a contractional regime (Korsch and Totterdell, 1995; Fig. 46). This is thought to have led to the formation of a retroforeland fold-thrust belt west of the magmatic arc, and the development of a major retroforeland basin phase in the Bowen-Gunnedah-Sydney basin system that continued until the Middle Triassic (Korsch and Totterdell, 1995). This period is, therefore, characterised by renewed arc magmatism, the foreland basin stage of development of the Bowen-Gunnedah-Sydney basin system and the Hunter-Bowen Orogeny (Figs 44, 45, 46).

The onset of the Hunter-Bowen Orogeny and the resultant change from extensional to contractional tectonism in the latest Early to mid Permian, is marked by a well-developed mid Permian unconformity in the backarc Bowen Basin and intracratonic Galilee and Cooper basins (e.g., Stephens et al., 1996; Korsch et al., 1998). The switch to contraction resulted in the transition of the Bowen-Gunnedah-Sydney basins from backarc extensional basins to foreland basins in a backarc setting (Roberts et al., 1996; Korsch et al., in press b).

Subsidence and sedimentation is suggested to have been driven by thrust loading related to westward-propagating thrust sheets from the New England Orogen (Korsch et al., in press b). The Bowen-Gunnedah-Sydney basin system retained this foreland basin setting until the Middle or Late Triassic (e.g., Harrington and Korsch, 1985; Korsch and Totterdell, 1995; Korsch et al., in press b). Foreland basin sedimentation is also recorded in the Gogango Thrust Zone, Yarrol Terrane, and in the Gympie Terrane (Fig. 44). Related forearc and accretionary wedge sedimentation in the New England Orogen is thought to be located further east (now offshore, e.g., Korsch, pers. comm., 2008.)

Terrestrial and marine sedimentation in the foreland (e.g., Bowen-Gunnedah-Sydney basins), and intracratonic basins (e.g., Cooper and Galilee basins), continued throughout the Permian and up to the Middle Triassic. Fluvial and lacustrine systems were associated with extensive peat swamps (coal measures) in the Cooper and Galilee basins (Gray and McKellar, 2002; Scott et al., 1995; Cowley, 2007; Fig. 44). In the Late Permian, coastal swamps also formed in the subsiding Bowen Basin, leading to an accumulation of extensive coal deposits (Shaw, 2002). Contemporaneous sedimentation is also recorded in the Parmeener Supergroup of the Tasmania Basin in Tasmania (Seymour and Calver, 1995, 1998). These contain an upper succession (Late Permian and younger) of nonmarine sediments, which also include coal measures (Seymour and Calver, 1995, 1998).

Permian glacial and marine sedimentation recorded (outcrop and sub-surface, e.g., beneath the Murray Basin) in Victoria and southern New South Wales do not appear to have continued into this cycle (O'Brien et al., 2003). Local, possibly originally more extensive, outcropping and concealed (by younger cover), Late Permian largely terrestrial sediments and coal measures occur in the Coen region and in the Hodgkinson Province (McConachie et al., 1997; Garrad and Bultitude, 1999; Fig. 44).

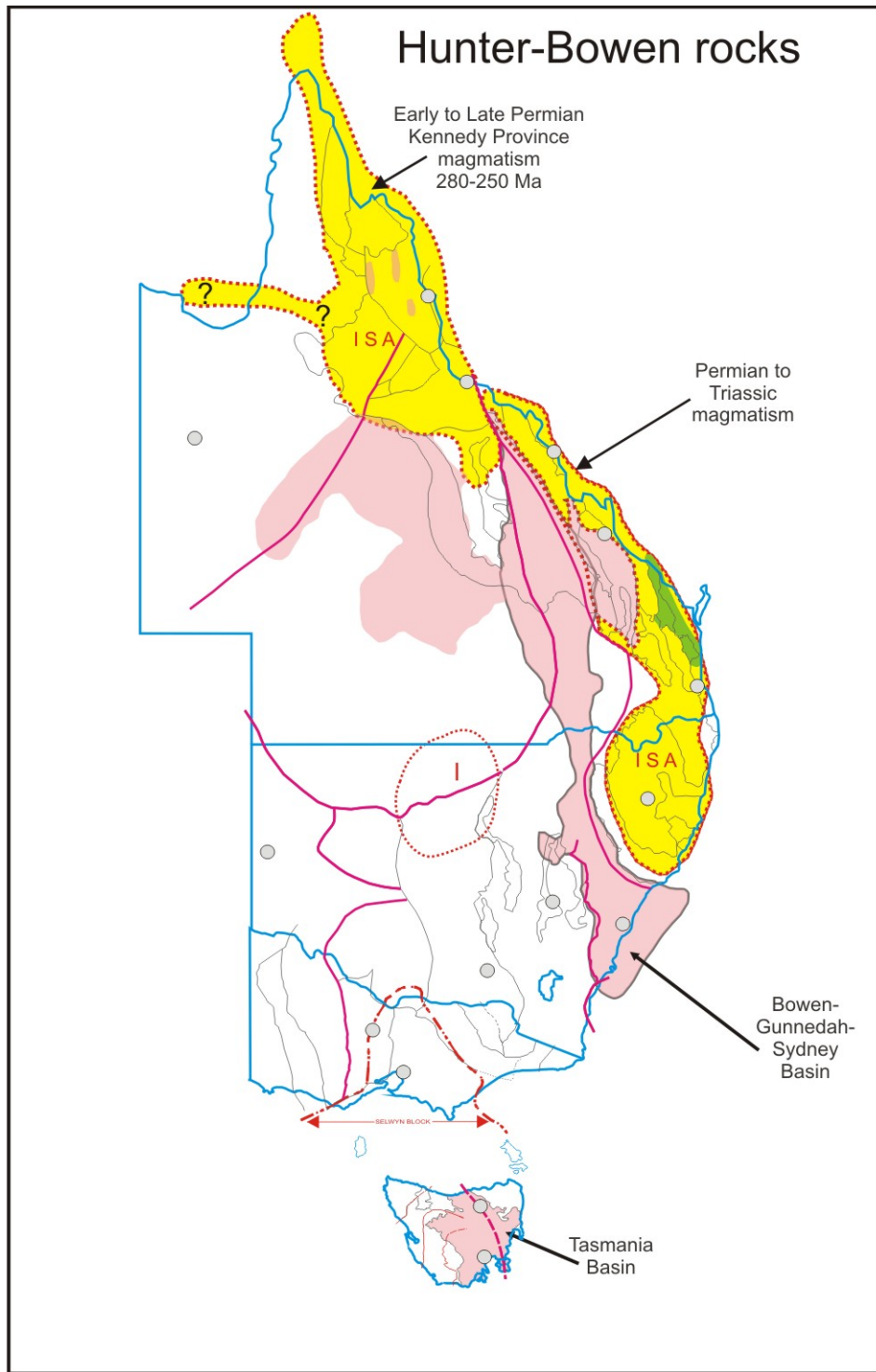


Figure 44. General distribution of early Late Permian to Middle-Late Triassic Hunter-Bowen cycle rocks in eastern Australia. Refer to text for discussion.

The late Early to early Late Permian saw a re-establishment of a magmatic arc along the Palaeo-Pacific continental margin of Australia, as recorded by voluminous Late Permian to Early (and Middle) Triassic calcalkaline, intrusive and extrusive, magmatism throughout the New England Orogen (ca. 270 Ma - 230 Ma; Gust et al., 1993; Bryant et al., 1997; Holcombe et al., 1997b; Van Noord, 1999; Murray, 2003; Figs 44, 46). Magmatism was dominantly andesitic to felsic in composition, consistent with a magmatic arc interpretation (Gust et al., 1993; Bryant et al., 1997; Holcombe et al., 1997b; Murray, 2003), although Murray (2003) notes that the granites are compositionally very similar to the earlier (Late Carboniferous to Early Permian), backarc, magmatism in the New England Orogen. Magmatism appears to wane in the Middle Triassic (Gust et al., 1993), although age control is poor. Widespread Middle to Late Permian intrusive and extrusive magmatism (and minor sediments) is also recorded in north Queensland, where it represents the youngest component of the voluminous Carboniferous-Permian Kennedy Province magmatism (Bain and Draper, 1997; Fig. 44). Late Permian Kennedy Province magmatism is largely confined to the eastern Hodgkinson Province (dominantly S-type; Bultitude and Champion, 1992; Bultitude et al., 1997; Garrad and Bultitude, 1999) and the Charters Towers region (I-type and mantle-derived; Hutton et al., 1997), but also includes A-, I- and mantle-derived extrusive and intrusive magmatism in the Coen, Georgetown and western Hodgkinson regions (Withnall et al., 1997a, Blewett et al., 1997; Fig. 44). It is generally thought to be broadly arc-related (e.g., Champion and Bultitude, 2003), probably in an extensional backarc or behind arc position relative to the continental arc in the New England Orogen (Fig. 46). Unlike the New England Orogen, magmatism in north Queensland ceased by the end of the Permian and no Early Triassic magmatism is recorded. West of the New England Orogen, restricted Triassic magmatic activity is recorded in the Cooper Basin (Murray, 1994; Draper, 2002a, b) and in the Bourke-Louth region (Burton et al., 2007; Fig. 44). These magmatic events suggest that tectonism on the eastern continental margin may have influenced the craton further west, for example, Burton et al. (2007) suggest that the Midway Granite, and coeval intrusives east of Bourke, indicate a more spatially widespread Middle Triassic magmatic pulse than is currently recognised.

The Hunter-Bowen Orogeny covers a period of about 35 m.y. from ~265 Ma to ~230 Ma, (Murray, 1997b; Holcombe et al., 1997b; Roberts et al., 2006; Korsch et al., in press b; Fig. 45) and encompasses all of the Hunter-Bowen cycle. The initiation of this event in the New England Orogen is marked by a major unconformity between early and late Permian rocks (Korsch et al., 1998). In the New England Orogen, deformation is characterised by retrothrusting driven by subduction further to the east. This major west-directed thrusting led to the formation of a retroforeland fold-thrust belt (Korsch et al., 1990; 1997; Fergusson, 1991; Holcombe et al., 1997; Korsch, 2004). Subsidence of the Bowen and Gunnedah basins during the foreland basin phase was driven by thrust loading related to these westward-propagating thrust sheets (Korsch et al., in press b). The foreland basin phase of sedimentation associated with the Hunter-Bowen Orogeny was punctuated by a series of discrete contractional events (Korsch et al., in press b) in the Permian and Triassic.

Recent detrital zircon age data suggest that the island arc component of the Gympie Terrane came into contact with the mainland New England Orogen, that is, continent-island arc collision, prior to ~250 Ma (Permian-Triassic boundary; Korsch et al., in press c). Approximately similar timing is recorded in north Queensland, where Late Permian east-west deformation, equated with the Hunter-Bowen Orogeny, is best developed in the Hodgkinson and Barnard provinces (Garrad and Bultitude, 1999; Fig. 45). Contractional events of the Hunter-Bowen Orogeny also appear to have thrust the Tamworth Belt westwards over the eastern edge of the Sydney-Gunnedah Basin and Lachlan Craton (Korsch et al., 1997; Roberts et al., 2004) around this time in the Late Permian and earliest Triassic.

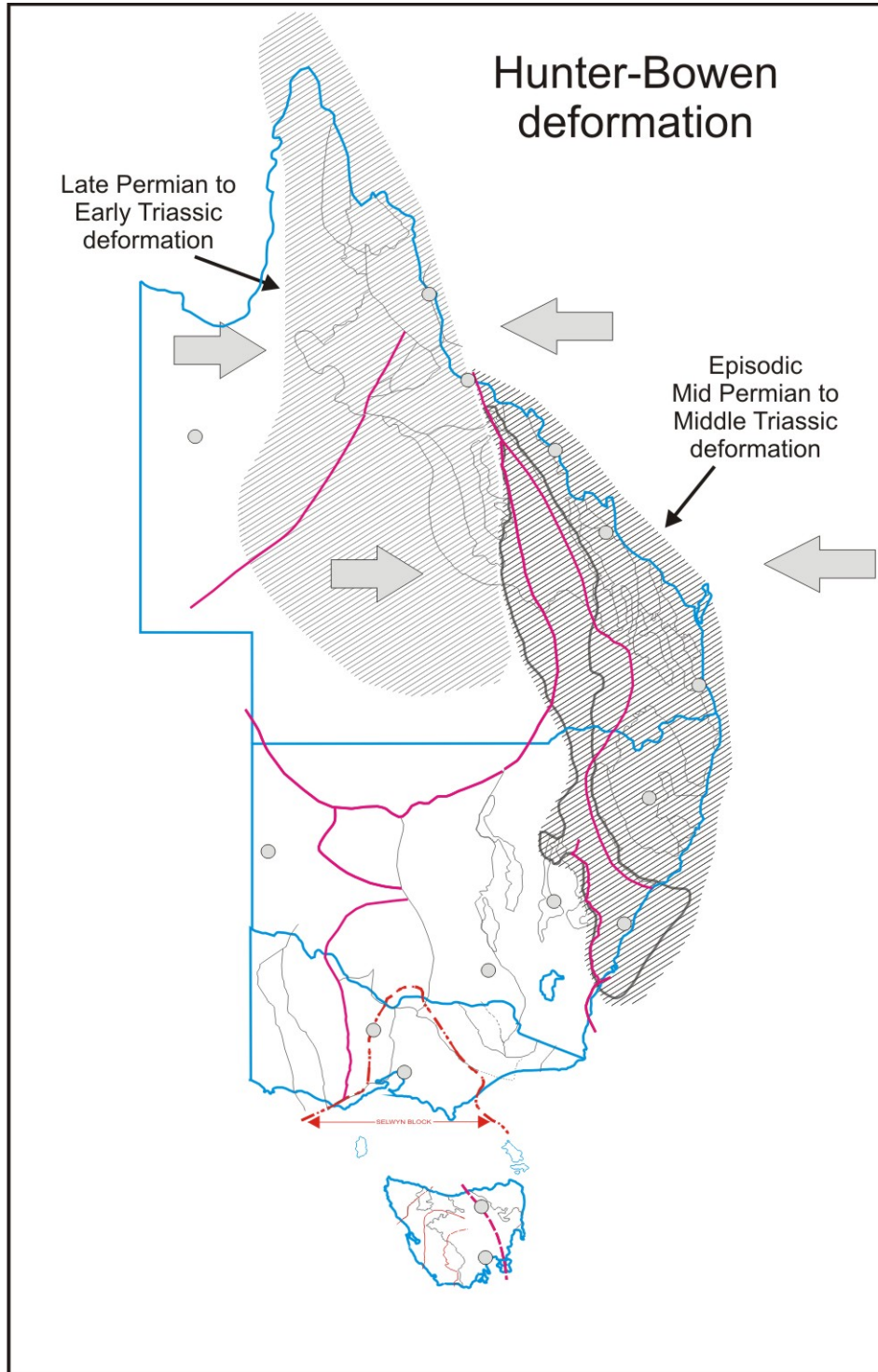


Figure 45. Generalised (approximate) distribution of deformation within the Hunter-Bowen Orogeny as defined by outcrop – actual extent of deformation may be greater and more continuous. The intensity of deformation is variable within the areas indicated. Deformation in the New England Orogen during this period was episodic, in apparent contrast in the region to the west. See text for data sources.

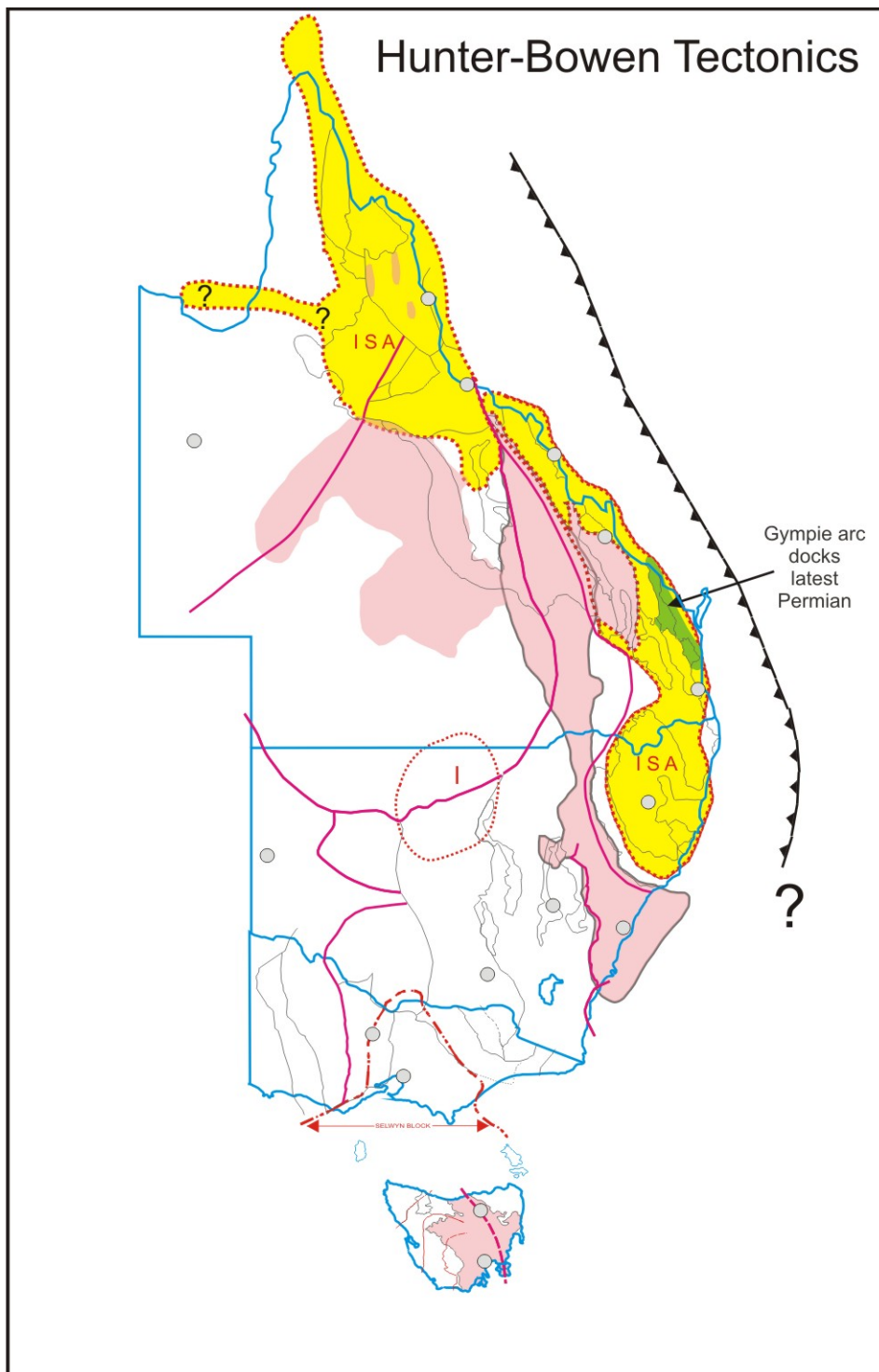


Figure 46. Interpreted tectonic model for the Hunter-Bowen cycle of eastern Australia. Refer to text for detailed discussion.

In the late Middle to Late Triassic, regional contraction resulted in uplift and erosion, and cessation of deposition in the Galilee, Cooper and Bowen basins (Apak et al., 1997; Bain and Draper, 1997; Green et al., 1997; Korsch et al., 1998), as well as east-west deformation and uplift in the Drummond Basin (Olgers, 1972; Fenton and Jackson, 1989; Murray, 1990; Johnson and Henderson, 1991; [Fig. 45](#)).

The Hunter-Bowen cycle marks the timing of effective cratonisation of eastern Australia. Following this cycle, there was a switch in geodynamics back to an extensional, probably backarc environment (e.g., Holcombe et al., 1997b). This resulted in a change in plutonism (A-type granites), felsic and/or bimodal volcanism (Stephens et al., 1993; Holcombe et al., 1997b) and development of extensional basins with coal-bearing successions (Holcombe et al., 1997b; Shaw, 2002). The change from backarc contraction and thrust loading to backarc extension was probably at ~230 Ma (R.J. Korsch, pers. comm. 2008).

PART 3. METALLOGENIC EVENTS IN PHANEROZOIC EASTERN AUSTRALIA

by DL Huston, N Kositcin and DC Champion

The previous two sections have provided an overview of the geology of the Tasman Orogen from the Late Neoproterozoic through to the Triassic and developed a framework for the geodynamic evolution of this region. In this section, we place metallogenic events into this framework and use these along with general relationships between metallogeny and geodynamics to predict mineral potential in eastern Australia. To provide an indication of this potential, basic data, including the age, size, geological characteristics and likely deposit type, are provided for major, or historically important, deposits and important prospects in the Tasman Orogen. [Table 1](#) summarises interpreted ages of formation for deposits that range in age between Late Neoproterozoic through Triassic. These data are presented in a space-time diagram in [Figure 47](#). The deposits are grouped by age into a series of mineralising events. Following the methodology of the previous two sections, descriptions of known Phanerozoic mineralising events in Australia are grouped according to the following time intervals:

1. Delamerian cycle: Late Neoproterozoic to Late Cambrian (600-490 Ma)
2. Benambran cycle: Late Cambrian to Earliest Silurian (490-430 Ma)
3. Tabberabberan cycle: Middle Silurian to Late Devonian (430-380 Ma)
4. Kanimblan cycle: Late Devonian to Late Devonian-Early Carboniferous (380-350 Ma)
5. Hunter-Bowen cycle: 350 Ma to 230 Ma

For each of these cycles, mineral potential analysis was undertaken using the distribution of known deposits and prospects and the predicted distribution of deposit groups based on the geodynamic framework described in Part 2. Using the pmd*CRC approach, this corresponds to questions 1 (geodynamic setting) and 2 (regional architecture). This mineral potential analysis is, in all cases, qualitative, and, in many cases, conceptual. We have not attempted quantitative mineral potential analysis.

3.1. Delamerian cycle (600-490 Ma)

Although relatively restricted in extent, Late Neoproterozoic to Late Cambrian rocks of the Tasman Orogen host a diverse suite of mineral deposits ([Fig. 47](#)). The earliest known mineral deposits in the Tasman Orogen are located along the western margin of the fold belt in association with Eocambrian to Middle Cambrian rift sequences in Western Tasmania and in the Koonenberry Belt in western New South Wales. These deposits include Ni-Cu and PGE deposits associated with gabbroic and ultramafic intrusive bodies. The Koonenberry Belt also contains a number of small Besshi-type VHMS deposits. However, the most significant deposits in this system are Kuroko-type VHMS and related deposits hosted by the Mount Read Volcanics of western Tasmania.

3.1.1. Ni-Cu deposits associated with the Crimson Creek Formation, western Tasmania (with contributions from R Bottrill)

Small gabbro bodies host small orthomagmatic Ni-Cu deposits (0.95 Mt at 0.76% Ni and 0.94% Cu: Seymour et al., 2006) in the Cuni field near Melba Flat ([Fig. 48](#)). Although there is some disagreement about age, we concur with Greenhill (1995) who interpreted these gabbros to have intruded correlates of the ~582 Ma (Calver et al., 2004; Seymour et al., 2006) Togari Group (specifically, the Crimson Creek Formation). Alternatively, the host sequence is part of the allochthonous Cleveland-Waratah assemblage (Brown, 1998). In either case, these deposits were emplaced prior to the 515-510 Ma Tyennan Orogeny, at which time mafic-ultramafic assemblages of the Waratah-Cleveland assemblage were obducted onto the east-facing Tasmanian passive margin (Crawford and Berry, 1992; Seymour et al., 2006). If the

interpretation of Greenhill (1995) is correct, the orthomagmatic Ni-Cu deposits of the Cuni field were emplaced during a phase of rift-related mafic volcanism that formed the Smithton Basin in northwest Tasmania and the Dundas Basin that underlies the Mount Read Volcanics in western Tasmania (Fig. 48).

Platinum group minerals (PGM) are widespread in Western Tasmania (particularly in the Heazlewood/Bald Hill, Adamsfield, Savage River and Wilson River areas: Bottrill, in prep). They were locally mined for a mixture of Os-Ir-Ru rich alloys known informally as “osmiridium”, but also containing some platinum group element sulphides and arsenides. Platinum, Pd, Rh and Au are locally important components. The PGM occur sparsely disseminated within the Cambrian Mafic-Ultramafic complexes, (e.g. at Adamsfield and Heazlewood), but most was produced from Cainozoic placer and residual deposits derived from the eroded ultramafic complexes. About 1 tonne was produced from about 1910-1960, about half being from Adamsfield and the rest mostly from the Heazlewood, Savage River and Wilson River areas.

The primary occurrence of the PGM is as sparsely disseminated grains within the ultramafic complexes, and are most commonly found within chromitite pods and layers, which may contain 5-10g/t of PGM. (Brown et al, 1988). They are also locally concentrated in some foliated serpentinites and talcose, limonitic joints or schlieren, as in Halls Open Cut at Adamsfield and Caudrys mine, Heazlewood (Reid, 1921; Nye, 1930). The latter mine produced about 250 oz of PGMs (Reid, 1921), while at Adamsfield, hard-rock production was estimated as between 200-400oz of PGMs (Nye, 1930). At Mt Stewart and Wilson River, osmiridium was found in schlieren transecting chromite and magnetite-rich serpentinite (Reid, 1921). These deposits are poorly understood.

At Adamsfield there has been significant osmiridium production from serpentine-rich quartzose sandstones and a cemented fragmental serpentinite overlying the serpentinite, and this appears to constitute an ancient shore-line placer of probable Ordovician age (Nye, 1930, Carey, 1952, Elliston, 1953). Western Tasmania has also produced nearly one tonne of osmiridium from placer districts spatially associated with the allochthonous Cleveland-Waratah assemblage (Brown, 1998). Although orthomagmatic PGE deposits have not been recognised in rocks of this assemblage, they are the most likely source of the placer deposits.

3.1.2. Ni-Cu and Zn-Pb deposits hosted by the Grey Range Group, Koonenberry Belt, western New South Wales

In the Koonenberry Belt, Gilmore et al. (2007) described both layered mafic-ultramafic hosted Ni-Cu and sediment-hosted Zn-Pb within the ~586 Ma Grey Range Group. The peridotite-hosted Mount Arrowsmith East Ni-Cu prospect has yielded intersections up to 0.50% Ni and 0.45% Cu associated with disseminated pyrite, chalcopyrite and pyrrhotite. The peridotite host rocks intrude Mount Arrowsmith Volcanics, which comprise alkali basalt, trachybasalt, trachyte and associated volcanoclastic rocks and subvolcanic intrusions (Gilmore et al., 2007). In addition, the laterally equivalent Kara Formation, which consists mostly of slates with lesser quartzite, dolomitic limestone, black shale, pyritic siltstone and exhalative units, has yielded historical intersections with anomalous Zn, Pb and Ag (Gilmore et al., 2007). Gilmore et al. (2007) interpreted the Grey Range Group as continental shelf deposits deposited during intracontinental rifting.

3.1.3. Cambrian mineralisation in the Koonenberry Belt, south-western New South Wales

Although the exact timing and style differ, both western Tasmania and the Koonenberry Belt are characterised by Middle to Late Cambrian VHMS deposits. In the Koonenberry Belt, the Ponto Group, which comprises distal continental shelf sediments associated with tholeiitic volcanics was deposited at ~510 Ma (Gilmore et al., 2007). As the Ponto Group is correlated with the calc-alkaline Mount Wright Volcanics, Gilmore et al. (2007) interpreted these rocks as forming in a back-arc basin with the arc located well to the east, or as a fore-arc basin with the Mount Wright Volcanics possibly representing the arc.

Table 1. Ages of selected mineral deposits in the Tasman Orogen, eastern Australia						
Deposit	Orogen	Deposit type	Type of age	Age (Ma)	Comments	Reference (s)
Delamerian cycle ((600-490 Ma)						
Grasmere	Delamerian	VHMS	Inferred	510	Age of host volcanic rocks.	Gilmore et al. (2007)
Hellyer	Delamerian	VHMS	Inferred	505	Age of host volcanic sequence.	Black et al. (1997)
Que Rive	Delamerian	VHMS	Inferred	505	Age of host volcanic sequence.	Black et al. (1997)
Rosebery	Delamerian	VHMS	Inferred	505	Age of host volcanic sequence.	Black et al. (1997)
Hercules	Delamerian	VHMS	Inferred	505	Age of host volcanic sequence.	Black et al. (1997)
Mt Lyell	Delamerian	Hybrid VHMS	Actual – ore mineral	500	Re-Os analysis of molybdenite.	D Huston, R Creaser and K Denwer (unpub. data)
Henty	Delamerian	Hybrid VHMS	Inferred	500	Inferred as same age as Mount Lyell.	
Benambran Cycle (490-430 Ma)						
Thalanga	Thomson-North Queensland	VHMS	Inferred	479	Age of host volcanic sequence.	Hutton et al. (1997)
Highway-Reward	Thomson-North Queensland	VHMS	Inferred	479	Age of host volcanic sequence.	Hutton et al. (1997)
Waterloo-Agincourt	Thomson-North Queensland	High sulphidation VHMS	Inferred	479	Age of host volcanic sequence.	Hutton et al. (1997)
Liontown	Thomson-North Queensland	VHMS	Inferred	479	Age of host volcanic sequence.	Hutton et al. (1997)
Balcooma	Thomson-North Queensland	VHMS	Inferred	478	Age of host volcanic rocks.	M Fanning in Rae (2000)
Dry River South	Thomson-North Queensland	VHMS	Inferred	478	Inferred as same age as Balcooma.	
Copper Hill	Lachlan	Porphyry Cu	Inferred	>447	K-Ar age of magmatic hornblende from ore-related diorite.	Perkins et al. (1995)
Marsden	Lachlan	Porphyry Cu	Inferred	447	Age of progenitor intrusion.	Crawford et al. (2007b)
Cadia	Lachlan	Porphyry Cu	Inferred	438-435	Age of progenitor intrusion.	Crawford et al. (2007b)
Northparkes (Goonumbla)	Lachlan	Porphyry Cu	Actual – alteration mineral	439	⁴⁰ Ar- ³⁹ Ar age of alteration minerals..	Perkins et al. (1995)
E39	Lachlan	Porphyry Cu	Actual – alteration mineral	440	⁴⁰ Ar- ³⁹ Ar age of alteration minerals.	Perkins et al. (1995)
Gidginbung	Lachlan	High sulphidation Cu-Au	Questionable (actual – ore mineral)	436	Age of inferred hydrothermal zircon; hydrothermal origin questioned by Fu et al. (2009).	Lawrie et al (2007)

Table 1. Ages of selected mineral deposits in the Tasman Orogen, eastern Australia						
Deposit	Orogen	Deposit type	Type of age	Age (Ma)	Comments	Reference (s)
Benambran Cycle (490-430 Ma)						
Gigdinbung	Lachlan	High sulphidation Cu-Au	Questionable (actual – alteration mineral)	417-401	^{40}Ar - ^{39}Ar age of alteration illite. Interpreted to have been reset.	Perkins et al. (1995)
Peak Hill	Lachlan	High sulphidation Cu-Au	Questionable	410	^{40}Ar - ^{39}Ar age of alteration minerals. Interpreted to have been reset. Preferred age is ~440 Ma.	Perkins et al. (1995)
Bendigo	Lachlan	Lode gold	Actual – alteration mineral	453; 441-435	^{40}Ar - ^{39}Ar age of alteration sericite.	Foster et al. (1998); Foster in Bierlein et al. (2001a)
Ballarat	Lachlan	Lode gold	Actual – alteration mineral	444-435	^{40}Ar - ^{39}Ar age of alteration sericite.	Bierlein et al. (1999)
Stawell – Magdala	Lachlan	Lode gold	Actual – alteration mineral	438	^{40}Ar - ^{39}Ar age of alteration sericite.	Foster et al. (1998)
Wattle gully	Lachlan	Lode gold	Actual – alteration mineral	441	^{40}Ar - ^{39}Ar age of alteration sericite.	Foster et al. (1998)
Tabberabberan cycle (430-380 Ma)						
Ardlethan	Lachlan	Intrusion-related Sn	Inferred	410	Rb-Sr age of inferred progenitor. Maximum age provided by ~417 Ma granite that hosted ore-bearing breccias.	Ren et al. (1995); Richards et al. (1982)
Kikoiria	Lachlan	Intrusion-related Sn	Inferred	430	Age of inferred progenitor granite.	Colquhoun et al. (2005)
Tara	Lachlan	Stockwork Cu-Zn-Pb-Ag	Actual – alteration mineral	420	^{40}Ar - ^{39}Ar age of vein-hosted muscovite.	Downes and Phillips (2006)
Mineral Hill	Lachlan	Epigenetic Au-Cu	Inferred	428-418	Age is constrained by age of oldest igneous rock in area and by age of unmineralised unit overlying deposit.	Blevin and Jones (2004); Morrison et al. (2004)
Holbrook	Lachlan	Intrusion-related Mo	Actual – ore mineral	423	Re-Os analysis of molybdenite.	Norman et al. (2004)
Lewis Ponds	Lachlan	VHMS	Inferred	417	Age of host rocks.	L Black in Huston et al. (1997)
Woodlawn	Lachlan	VHMS	Inferred	419	Age of host rocks.	Bodorkos and Simpson (2008)
Lake George (Captains Flat)	Lachlan	VHMS	Inferred	420	Correlation with unit hosting Lewis Ponds deposit.	
Currawong-Wilga	Lachlan	VHMS	Inferred	420	Correlation with unit hosting Lewis Ponds deposit.	

Table 1. Ages of selected mineral deposits in the Tasman Orogen, eastern Australia						
Deposit	Orogen	Deposit type	Type of age	Age (Ma)	Comments	Reference (s)
Tabberabberan cycle (430-380 Ma)						
Kempfield	Lachlan	VHMS – barite	Inferred	420	Correlation with unit hosting Lewis Ponds deposit.	
Dargues Reef (Majors Creek)	Lachlan	Intrusion-related Au	Actual – alteration mineral	411-401	K-Ar dating of ore-related sericite. Similar to age of host Braidwood Granodiorite	McQueen and Perkins (1995; Bodorkos and Simpson (2008)
Tarnagulla	Lachlan	Lode Au	Actual – alteration mineral	410	⁴⁰ Ar- ³⁹ Ar age of ore-related muscovite.	Bierlein et al. (2001a)
Stawell – Wonga	Lachlan	Lode Au	Inferred	423-400	Constrained by age of dykes hosting of and age of granite that has thermally metamorphosed the ores	Wilson et al. (1999); Phillips et al. (2003)
Charters Towers	Thomson-North Queensland	Lode Au	Actual – alteration mineral	412-400	⁴⁰ Ar- ³⁹ Ar age of alteration sericite. Average is 407 Ma.	Kruezer (2005)
The Peak	Lachlan	Structurally controlled Cu-Au	Actual – alteration mineral	405-400	⁴⁰ Ar- ³⁹ Ar age of alteration muscovite.	Glen et al. (1992); Perkins et al. (1994)
CSA	Lachlan	Structurally controlled Cu-Au	Actual – alteration mineral	405-400	⁴⁰ Ar- ³⁹ Ar age of alteration muscovite.	Glen et al. (1992); Perkins et al. (1994)
Beaconsfield	Delamerian	Lode Au	Actual – alteration mineral	400	⁴⁰ Ar- ³⁹ Ar age of alteration fuchsite.	Bierlein et al. (2005)
Mathinna-Mangana	Lachlan	Lode Au	Actual – alteration mineral	394-390	⁴⁰ Ar- ³⁹ Ar age of alteration muscovite.	Bierlein et al. (2005)
Adelong	Lachlan	Lode Au	Actual – alteration mineral	393-389	⁴⁰ Ar- ³⁹ Ar age of alteration sericite.	Perkins et al. (1995)
Lisle-Golconda	Lachlan	Lode Au	Actual – alteration mineral	385	⁴⁰ Ar- ³⁹ Ar age of alteration muscovite.	Bierlein et al. (2005)
Endeavour (Elura)	Lachlan	Structurally controlled Zn-Pb-Ag	Actual – alteration mineral	384	⁴⁰ Ar- ³⁹ Ar age of alteration muscovite.	Glen et al. (1992); Perkins et al. (1994)
King River	Lachlan	Lode Au	Actual – alteration assemblage	383	⁴⁰ Ar- ³⁹ Ar age of silicified pelite.	Bierlein et al. (2005)
Mount Morgan	Thomson-North Queensland	Hybrid VHMS	Inferred	>381	Age of Mount Morgan Tonalite that has contact metamorphosed the ores.	Golding et al. (1993)
Kanimblan cycle (380-350 Ma)						
Woods Point goldfield	Lachlan	Lode Au	Actual – alteration mineral	378-372	⁴⁰ Ar- ³⁹ Ar age of alteration sericite.	Foster et al. (1998)

Table 1. Ages of selected mineral deposits in the Tasman Orogen, eastern Australia						
Deposit	Orogen	Deposit type	Type of age	Age (Ma)	Comments	Reference (s)
Kanimblan cycle (380-350 Ma)						
Fosterville	Lachlan	Lode Au	Actual – alteration mineral	381	⁴⁰ Ar- ³⁹ Ar age of altered rhyolite dyke.	Bierlein et al. (2001a)
Maldon – Day Dawn	Lachlan	Intrusion-related Au (?)	Maximum	>370	Age of diorite dyke hosting mineralisation. Considered most likely age of mineralisation.	Bierlein et al. (2001a)
Anchor	Lachlan	Intrusion-related Sn (greisen)	Inferred	378	Age of inferred progenitor Lottah Granite	Black et al. (2005)
Aberfoyle and Story's Creek	Lachlan	Intrusion-related Sn-W (vein)	Inferred	377	Age of inferred progenitor Henbury Granite	Black et al. (2005)
Queen Hill	Delamerian	Intrusion-related Sn (carbonate replacement)	Inferred	361	Age of inferred progenitor Heemskirk Granite	Black et al. (2005)
Hill End	Lachlan	Lode Au	Actual – alteration mineral	358	⁴⁰ Ar- ³⁹ Ar age of alteration muscovite.	Lu et al. (1996)
Bold Head-Dolphin	Delamerian	Intrusion-related W (skarn)	Inferred	351	Age of inferred progenitor Heemskirk Granite	Black et al. (2005)
Hunter-Bowen Cycle (350-230 Ma)						
Herberton Sn-W province	Thomson-North Queensland	Intrusion-related Sn-W	Inferred	315	Age of inferred progenitor O'Briens Creek Supersuite	Geological Survey of Queensland, unpublished data
Cooktown Sn province	Thomson-North Queensland	Intrusion-related Sn	Inferred	260	Age of inferred progenitor Cooktown Supersuite	Bultitude and Champion (1992)
Mount Carbine Sn-W province	Thomson-North Queensland	Intrusion-related Sn-W	Inferred	280	Age of inferred progenitor Wypalla Supersuite	Bultitude and Champion (1992)
Hill End	Lachlan	Lode Au	Actual – alteration mineral	343	⁴⁰ Ar- ³⁹ Ar age of alteration muscovite.	Lu et al. (1996)
Pajingo (Scott)	Thomson-North Queensland	Low sulphidation epithermal Au-Ag	Actual – alteration mineral	342	K-Ar age of alteration sericite	Richards et al. (1998)
Kidston	Thomson-North Queensland	Intrusion-related Au	Actual – alteration mineral	332	⁴⁰ Ar- ³⁹ Ar age of alteration sericite. Confirmed by age of syn- and post-mineralisation dykes.	Perkins and Kennedy (1998)

Table 1. Ages of selected mineral deposits in the Tasman Orogen, eastern Australia						
Deposit	Orogen	Deposit type	Type of age	Age (Ma)	Comments	Reference (s)
Hunter-Bowen Cycle (350-230 Ma)						
Ravenswood	Thomson-North Queensland	Intrusion-related Au	Actual – alteration mineral	330-310	^{40}Ar - ^{39}Ar ages of alteration sericite and biotite.	Perkins and Kennedy (1998)
Red Dome-Mungana	Thomson-North Queensland	Intrusion-related Au	Inferred	320-308	Age range defined by age of inferred progenitor porphyry dyke (upper) and ^{40}Ar - ^{39}Ar age of alteration sericite (lower).	Perkins and Kennedy (1998)
Mount Wright	Thomson-North Queensland	Intrusion-related Au	Actual – alteration mineral	306	^{40}Ar - ^{39}Ar age of alteration sericite.	Perkins and Kennedy (1998)
Cracow	New England	Low sulphidation epithermal Au-Ag	Inferred	291	Age of rhyolite dyke inferred to be ore-related.	C Perkins in Dong and Zhou (1996)
Mount Leyshon	Thomson-North Queensland	Intrusion-related Au	Actual – alteration mineral	290	^{40}Ar - ^{39}Ar age of alteration sericite.	Perkins and Kennedy (1998)
Mount Chalmers	New England	Hybrid VHMS	Inferred	277	Age of host volcanics.	Crouch (1999)
Hilgrove	New England	Lode Sb-Au	Inferred	255	K-Ar age of phlogopite from ore-related lamprophyre dykes.	Ashley et al. (1994)
Timbarra	New England	Intrusion-related Au	Actual – ore and alteration minerals; inferred	248-246	K-Ar, ^{40}Ar - ^{39}Ar and Rb-Sr ages of alteration minerals; ages of magmatic zircon and monazite and of hydrothermal xenotime.	Kleeman et al (1997); Schaltegger et al. (2005); Pettke et al. (2005)
Gympie	New England	Low sulphidation epithermal Au-Ag	Actual – alteration mineral	245	Age of hydrothermal sericite; method not described.	Cuneen (1996)
Ruby Creek mineral field	New England	Intrusion-related Sn-W-Mo-Cu-Bi (veins, pipes and greisens)	Actual – ore mineral	242	Re-Os age of molybdenite.	Norman et al. (2004)
Doradilla	Lachlan	Intrusion-related Sn	Inferred	231	Age of felsic dykes associated with deposits.	Burton et al. (2007)

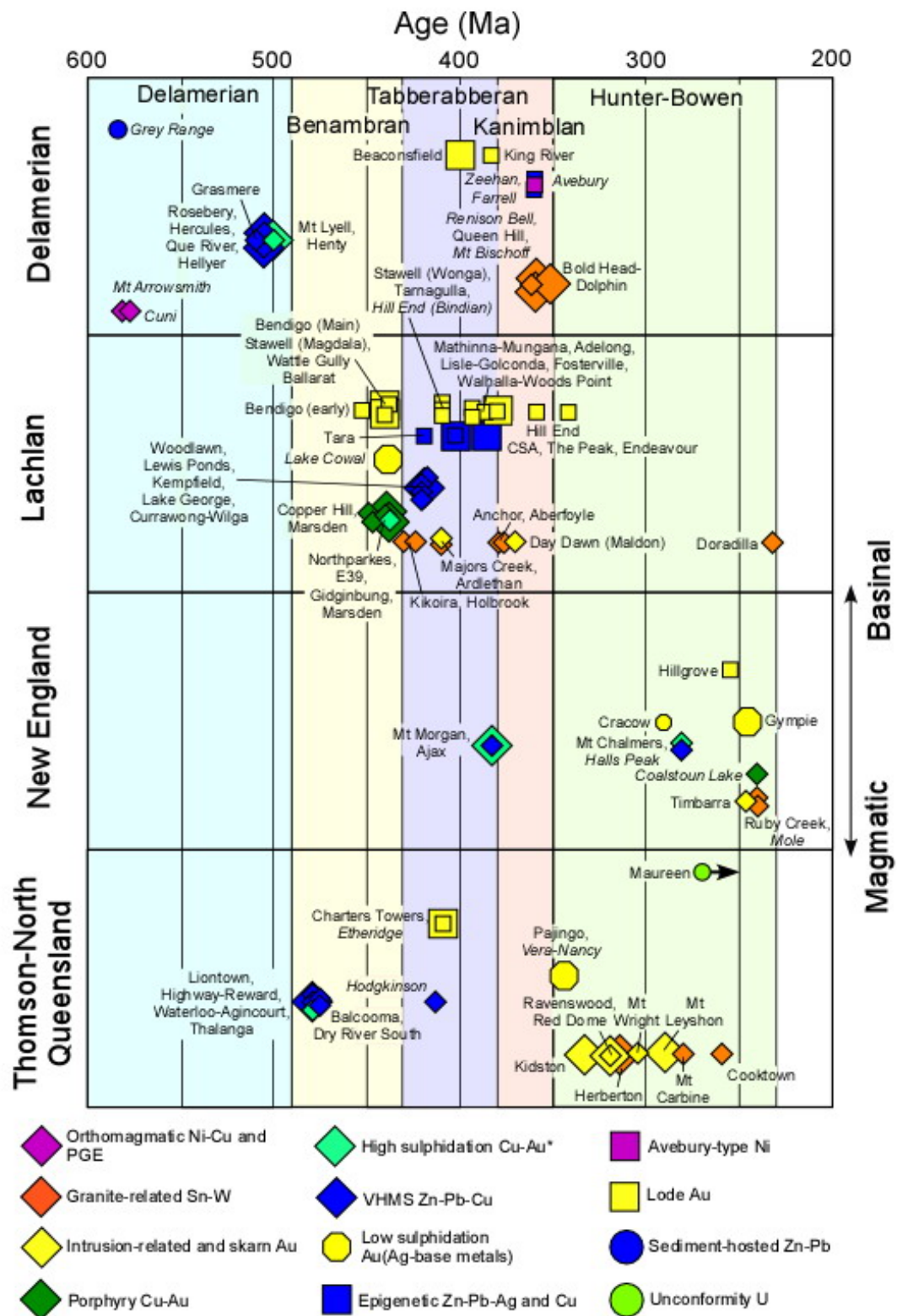


Figure 47. Space-time diagram showing the ages of selected mineral deposits in the Lachlan Orogen, eastern Australia.

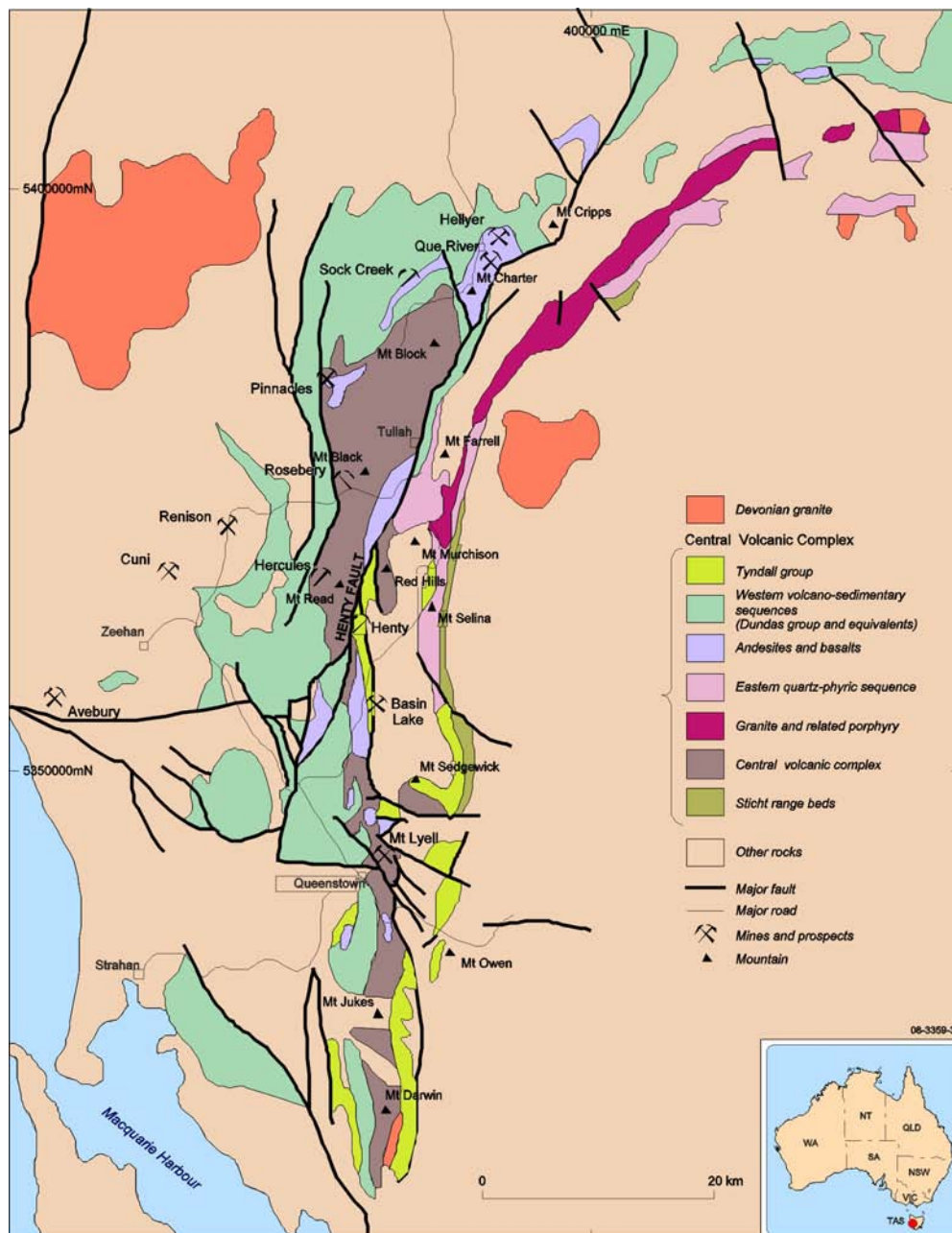


Figure 48. Geology of western Tasmania, showing the location of important deposits and prospects.

The Ponto Group contains a number of small polymetallic deposits, some of which were mined around the turn of the 20th century. These include the Grasmere deposit, which has a JORC-compliant resource of 0.584 Mt grading 2.47% Cu and 0.94% Zn, as well as the Ponto and the Cymbric Vale prospects. Gilmore et al. (2007) interpreted these deposits as Besshi-type VHMS deposits.

3.1.4. VHMS and related deposits in the Mount Read Volcanics, western Tasmania

The Cambrian Mount Read volcanic belt in western Tasmania (Fig. 48) is Australia's most significant VHMS province, with pre-mining global resources of 8.1 Mt Zn, 3.0 Mt Pb, 3.3 Mt Cu, 9.1 kt Ag and 278 tonnes Au (Seymour et al., 2006). These deposits can be split into two groups according to their metallogeny: Zn-Pb and Cu-Au. Differences between these two groups extend beyond metallogeny to include differences in the form and style of the ore, in associated alteration assemblages, in sulphur and Pb isotope systematics, and, possibly, in age.

In addition to having low Cu/(Cu+Zn) ratios, deposits of the Zn-Pb group, which includes the Hellyer, Que River, Rosebery, Hercules, South Hercules, Mount Charter and Tasman/Crown Lyell deposits, are characterised by massive, stratiform ores that are dominated by pyrite, sphalerite, galena and barite and are confined to breaks in volcanic activity, as marked by changes of volcanic composition and/or fine-grained siliciclastic rocks (Fig. 49a,b). These deposits are interpreted to have formed either at or just below the seafloor in response to the mixing of upwelling ore fluids with cold seawater (Green et al., 1981; Gemmell and Large, 1992). The alteration zones associated with these deposits are dominated by quartz-chlorite±carbonate (e.g., Hellyer: Gemmell and Large, 1992; Gemmell and Fulton, 2001) and quartz-sericite-pyrite (e.g., Rosebery: Green et al., 1981; Large et al., 2001) assemblages. The form and associated alteration assemblages of these Zn-Pb-rich deposits are typical of "normal" VHMS deposits worldwide. With the exception of the Tasman/Crown Lyell deposits, which are hosted by the overlying Tyndall Group, these syngenetic deposits are hosted in the Central Volcanic Sequence, which has an age of ~505 Ma (Black et al., 1997).

In contrast, the Cu-Au group of deposits, which include those in the Mount Lyell field, and the Garfield, Basin Lake and Henty/Mount Julia deposits (Fig. 48), are characterised mostly by disseminated, stratabound ores that have replaced volcanic rocks. Important exceptions to this generalisation are the Blow and South Lyell deposits, the only stratiform Cu-Au-rich massive sulphide bodies in the Mount Lyell field. The most striking differences between the Cu-Au group and the Zn-Pb group are the ore and, particularly, alteration assemblages. Although most deposits of the Cu-Au group are dominated by a chalcopyrite-pyrite ore assemblage, zones within many orebodies contain significant bornite that is associated with minor to trace chalcocite, mawsonite, molybdenite, hematite, enargite, barite and wolframite (Walshe and Solomon, 1981; Manning, 1990; Huston and Kamprad, 2001; D Huston, unpub. data). As first recognised by Cox (1981), many of the deposits in the Mount Lyell field contain significant pyrophyllite and topaz in alteration assemblages. After documenting the zonation of pyrophyllite and topaz (Fig. 49c) and recognising the presence of woodhouseite and zunyite at the Western Tharsis deposit, Huston and Kamprad (2001) suggested the minerals were part of an advanced argillic alteration assemblage and that the Mount Lyell field was in part a high sulphidation Cu-Au system. Since then, studies of the deeper parts of the Henty-Mount Julia Au system have documented similar advanced argillic alteration assemblages (Callaghan, 2001; Halley, 2008), and K Denwer (pers. comm., 2007) has documented similar alteration zonation in other deposits in the Mount Lyell field, suggesting that the Mount Lyell field and the Henty-Mount Julia deposit are expressions of the same high sulphidation hydrothermal system, with the latter having formed in the upper levels of the system. The Basin Lake system, midway between the Henty deposit and the Mount Lyell field, is also characterised by advanced argillic assemblages (Williams and Davidson, 2004). These deposits are hosted either in the upper part of the Central Volcanic Complex or in the Tyndall Group. This, and the likelihood that a large proportion of the ores are replacive, suggests that these Cu-Au deposits are somewhat younger than the ~505 Ma Zn-Pb group of deposits. Re-Os dating of molybdenite from the Prince Lyell deposit in the Mount Lyell field indicates an age of 500.4 ± 2.3 Ma (2σ ; D Huston, R Creaser and K Denwer, unpub. data). Geological and isotopic assemblages point to major genetic differences between Zn-Pb and Cu-Au group deposits in the Mount Read Volcanics. The forms, metal associations and alteration assemblages of Zn-Pb group deposits are fairly typical of "normal" VHMS deposits worldwide. In contrast, the alteration assemblages associated with the Cu-Au group of deposits are not typical of VHMS assemblages, having more in common with advanced argillic assemblages associated with high sulphidation epithermal deposits. These assemblages suggest a major magmatic-hydrothermal contribution to the Mount Lyell ores, consistent with the results of Large et al. (1996), who suggested a

magmatic-hydrothermal component to the Mount Lyell system based on the presence of magnetite-apatite assemblages in the ores and a similar assemblage spatially associated with Cambrian granites.

3.1.5. Mineral potential

Similarities in age (~580 Ma), lithologic assemblages (e.g., mafic (tholeiitic) volcanic rocks combined with slate-shale-dolomite assemblages), and tectonic setting (rift) indicate linked genesis of the Dundas and Smithton Troughs in Tasmania with the Koonenberry Belt, particularly as this similarity extends to younger periods (510-505 Ma: see below). If this hypothesis is true, it suggests that potential for orthomagmatic Ni-Cu-PGE deposits may exist in areas between western Tasmania and the Koonenberry Belt in a similar position relative to older rocks of the North and South Australian Cratons to the west (Fig. 50). This area of potential includes the extension of the Koonenberry Belt to the south under the Murray-Darling Basin and the Stavely Belt in western Victoria.

In Tasmania, the association of osmiridium placers with exposures of the Cleveland-Waratah assemblage suggests that this assemblage has significant unrealised potential for hosting orthomagmatic PGE(Ni-Cu) deposits, in particular, the lower Cambrian ultramafic rocks for Os-Ir-Ru and Ni deposits. Brown (1992) reported the presence of pentlandite and PGE minerals in the Serpentine Hill Ultramafic Complex, which forms part of the Cleveland-Waratah assemblage. 3D geological models of Tasmania (Seymour et al., 2006) suggest that allochthonous ultramafic-mafic complexes of the Cleveland-Waratah assemblage, which were obducted during the 515-510 Ma Tyennan Orogeny (Crawford and Berry, 1992; Seymour et al., 2006), are widespread throughout much of Tasmania, particularly the Western Tasmanian Terrane. In addition, gabbroic rocks coeval with the Togari Group also have potential for Ni-Cu-Pt-Pd-Au deposits.

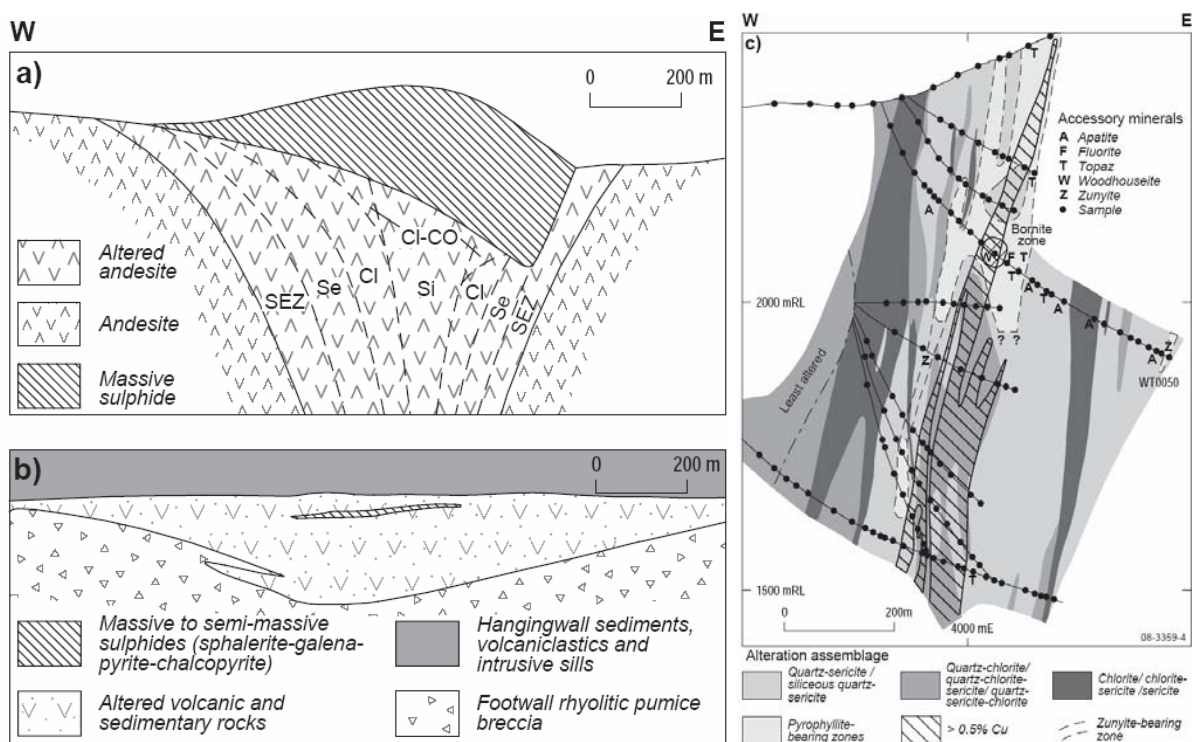


Figure 49. Cross sections of the (A) Rosebery, (B) Hellyer, and (C) Western Tharsis deposits in western Tasmania.

During the Middle to Late Cambrian, the greatest potential in the Tasman Orogen is in the Mount Read Volcanics for Kuroko-type VHMS and related deposits, although time equivalent rocks in the Koonenberry and Stavely Belts (Fig. 50), the felsic volcanic rocks in the Anakie Inlier also have potential for VHMS and related deposits, particularly of the Besshi-type (e.g., Peak Downs in the Anakie Inlier). High sulfidation,

VHMS-related Cu-Au deposits such as in the Mount Lyell field and Henty in Tasmania appear to be concentrated on the eastern margin of the Mount Read Volcanic Belt, more proximal to the possible position of an associated volcanic arc, if indeed the volcanism is backarc related (Crawford and Berry, 1992 interpreted the setting as post-collisional). Hence, Cu-Au-rich deposits are more likely to be located in arc-proximal (or rifted arc) positions within back-arc basins, as has been suggested for the Doyon-Bousquet-LaRonde district in the Abitibi Subprovince (Mercier-Langevin et al., 2007). In contrast, Zn-rich Kuroko-type deposits may be located more distal from the arcs in back-arc basins. However, both the Cu-Au-rich and Zn-rich deposits are most likely to be associated with volcanic centres. The possibility that the deposits hosted by the Ponto Group are hosted in a fore-arc basin also raises the possibility for analogous geodynamic settings of this and younger ages (see below) to host VHMS deposits. In addition to potential for VHMS and related deposits, mafic-ultramafic complexes have potential for hydrothermal Ni deposits where they are intruded by Delamerian cycle granites, possibly in western Tasmania (e.g., Cleveland-Waratah assemblage) and the Koonenberry Belt.

In addition to potential already described for the Mount Read-Stavely-Koonenberry Belt, we consider that other Delamerian cycle rocks, including extensions below the Murray-Darling Basin, have some potential for Cu-Au, Ni-Cu and Sn deposits. For example, the dismembered Jamieson-Licola belt in eastern Victoria, which Gray and Foster (2004) interpreted as an island arc (c.f. Vandenberg et al., 2000), has potential for hybrid and/or porphyry Cu-Au deposits. Similarly, Cambrian mafic-ultramafic complexes in the Dimboola Igneous Complex of central Victoria have potential for orthomagmatic Ni-Cu deposits, and felsic volcanic rocks in the basal Warburton Basin, along the western margin of the Thomson Orogen, have potential for VHMS and porphyry or hybrid Cu-Au deposits (Fig. 50). Island arc and oceanic floor remnants, with ages that range from the Delamerian through the Tabberabberan cycles, along the Peel-Manning Fault in northeastern New South Wales, have potential for porphyry Cu, VHMS and orthomagmatic Cu-Ni deposits. Delamerian cycle granites in southwestern Victoria (I-type) and in the Anakie Inlier (S-type) have potential for associated Sn and/or W deposits.

Rocks older than 490 Ma that have been affected by the Delamerian Orogeny (Fig. 51) may have potential for lode Au deposits. All other major orogenies, including the Benambran, Tabberabberan and Kanimblan Orogenies, in the Tasman Orogen have temporally and spatially associated lode Au events (see below), raising the expectation for a similar association in Delamerian-age fold belts. In addition, Cobar-type structurally controlled Cu-Au and Zn-Pb deposits may be associated with Delamerian inversion of Delamerian cycle deep water basins, particularly in the Koonenberry Belt and in the Thomson Orogen.

3.2. Benambran cycle (490-430 Ma)

The period between 490 and 430 Ma is one of the most richly mineralised periods (Fig. 47) in the geologic history of eastern Australia, including 455-440 Ma lode Au deposits that account for the vast majority of Au in the world-class Victorian goldfield, significant 450-440 Ma porphyry Cu-Au and related epithermal Au deposits in the Macquarie Arc of central New South Wales, and moderate-sized VHMS deposits in the ~480 Ma Seventy Mile Range Group, Balcooma Metavolcanic Group and Eland Metavolcanics in northern Queensland. In addition, a minor Irish-style Zn-Pb event occurred in Tasmania at ~445 Ma. The coincidence in time between the Macquarie porphyry-epithermal and the Victoria lode Au systems may suggest that these two systems were responses to a single geodynamic system.

3.2.1. VHMS and related deposits, Seventy Mile Range Group and Balcooma Metamorphics

Early Ordovician rocks of the Seventy Mile Range Group, Balcooma Metavolcanic Group and Eland Metavolcanics form two semi-continuous belts along the northern margin of the Thomson Orogen in northern Queensland (Fig. 52). The Seventy Mile Range Group consists of mostly low metamorphic grade volcanic and related sedimentary rocks that form an east-west trending belt that can be traced for over 200 km to the south of Charters Towers. The Balcooma Metavolcanic Group and Eland Metavolcanics form a

north-northeast trending belt of medium metamorphic grade metavolcanic and metasedimentary rocks that can be traced discontinuously over 100 km to the southeast of Einasleigh. Although these two belts of rocks are now separated over 200 km, lithological similarities and limited geochronological data suggest that they were once continuous. Both belts contain VHMS deposits, with global resources of 1.0 Mt Zn, 0.3Mt Pb, 0.37 Mt Cu, 3.5 t Au and 0.72 kt Ag, and 0.35 Mt Zn, 0.15 Mt Pb, 0.12 Mt Cu, 3.9 t Au and 0.36 kt Ag for the Seventy Mile Range and Balcooma belts, respectively (c.f. Hutton and Withnall, 2007).

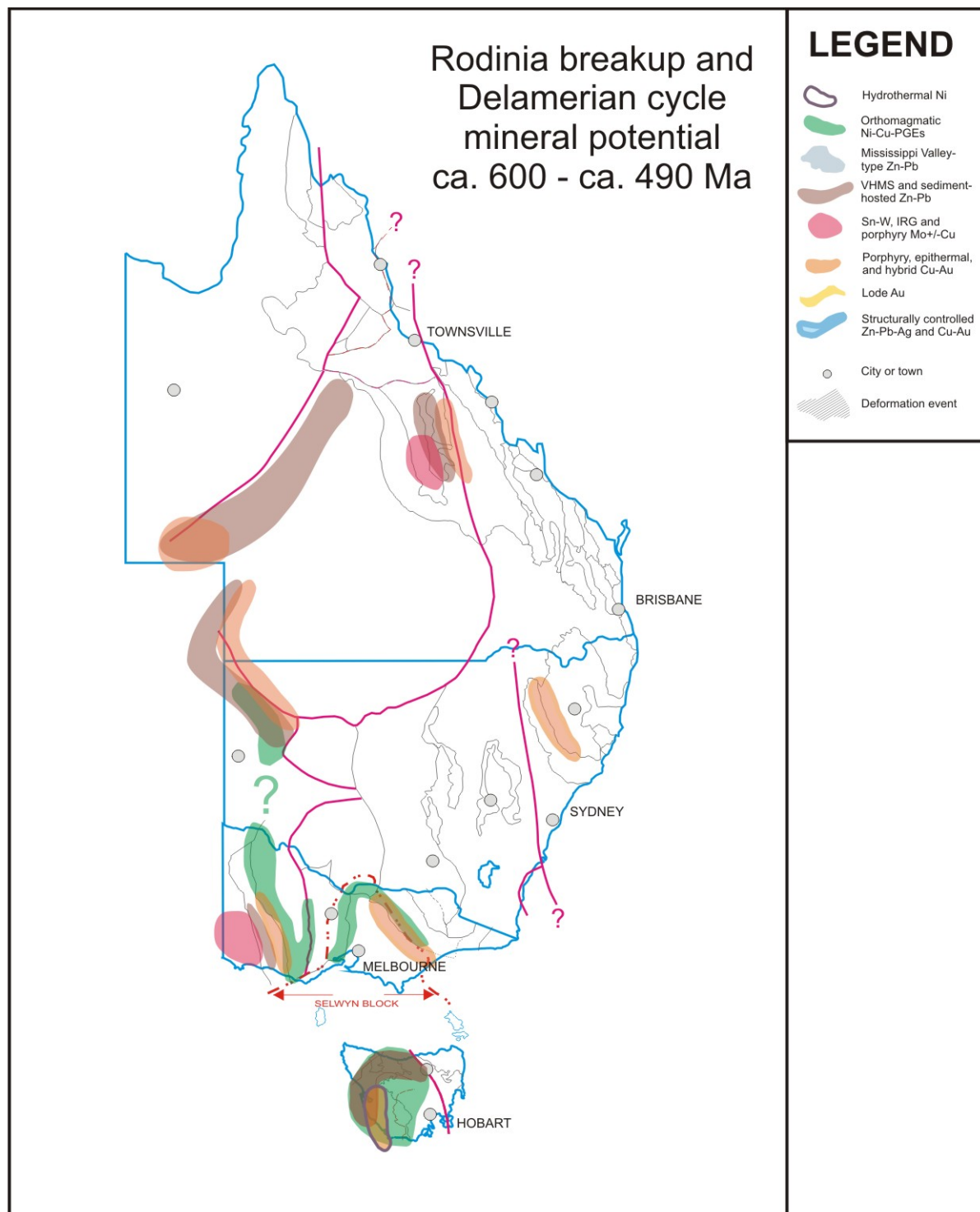


Figure 50. Mineral potential of the Delamerian cycle.

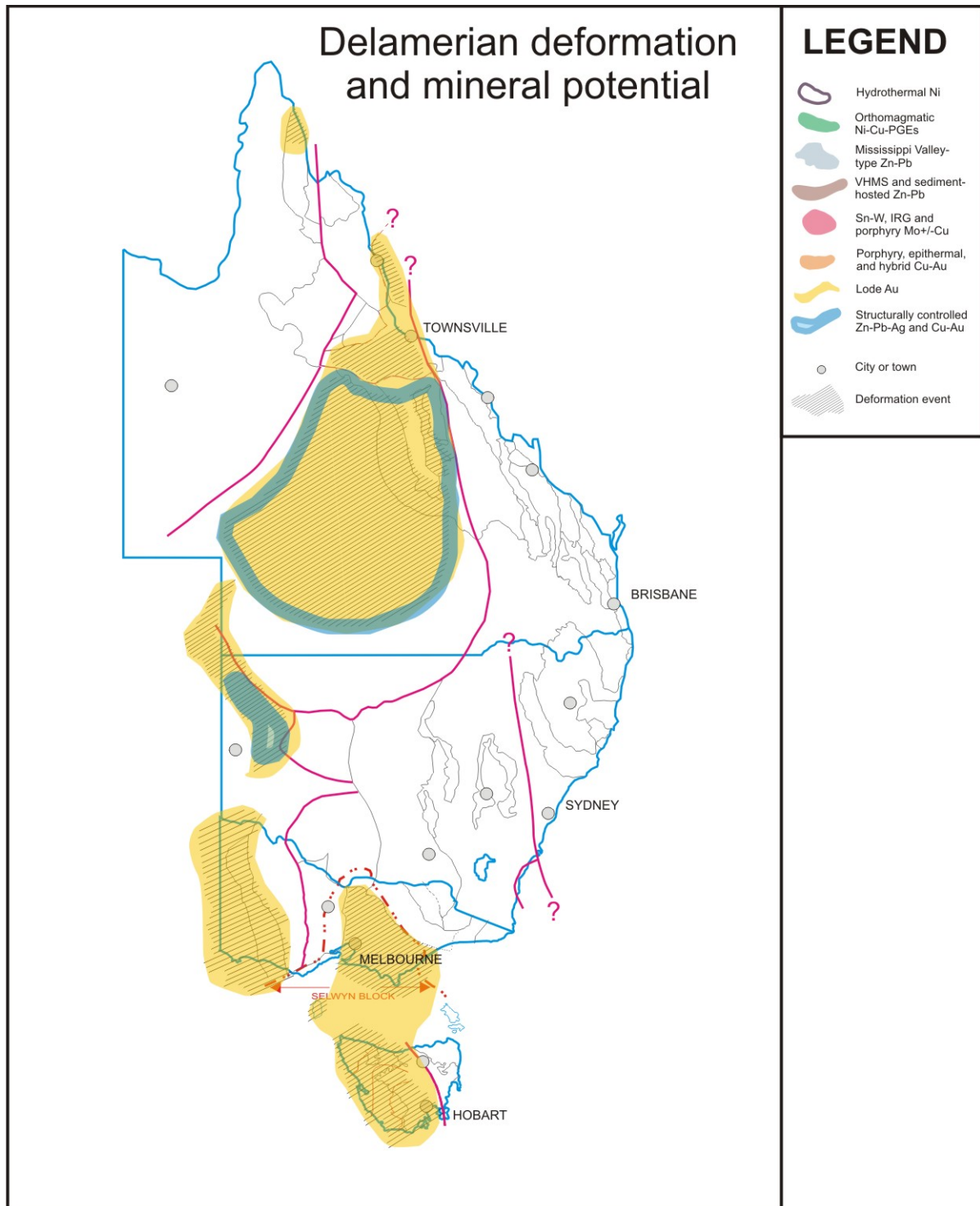


Figure 51. Mineral potential of the Delamerian Orogeny.

The Seventy Mile Range Group (Fig. 52) comprises four units, the Puddler Creek Formation, the Mount Windsor Volcanics, the Trooper Creek Formation, and the Rollston Range Formation (Henderson, 1986; Berry et al., 1992). The Puddler Creek Formation comprises continentally derived siliciclastic rocks intruded by mafic dykes. The Mount Windsor Volcanics are dominated by rhyolitic to dacitic volcanic rocks with minor andesite, whereas the overlying Trooper Creek Formation is dominated by intermediate to mafic volcanism with associated siliciclastics. The uppermost Rollston Range Formation comprises volcanoclastic sandstone and siltstone (Henderson, 1986; Berry et al., 1992; Hutton and Withnall, 2007). An

unpublished age of ~479 Ma for the Mount Windsor Volcanics implies a lower Ordovician age for this sequence (Hutton et al., 1997).

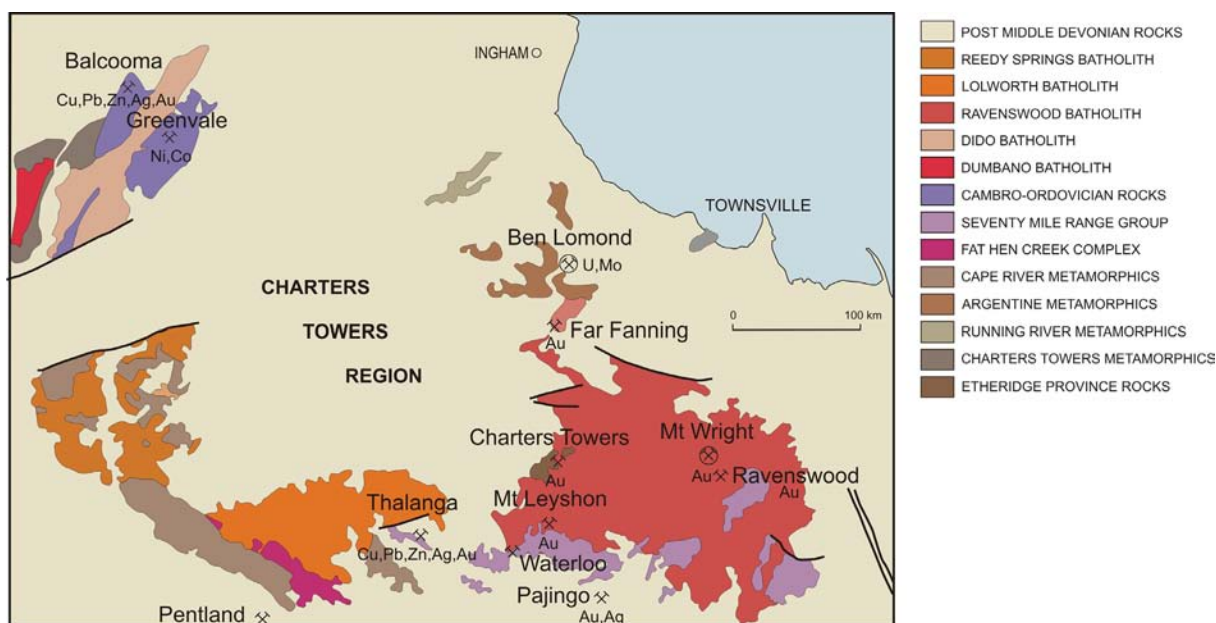


Figure 52. Geology of Thalanga Province, north Queensland, showing locations of Ordovician (meta-)volcanic belts and VHMS deposits and prospects.

The major deposit in the Seventy Mile Range Group, Thalanga (6.35 Mt @12.3% Zn, 3.9% Pb, 2.2% Cu and 99 g/t Ag; Hutton and Withnall, 2007), is localised along the contact between the Mount Windsor Volcanics and the Trooper Creek Formation. This deposit is tabular and is associated with a quartz-sericite-pyrite±chlorite alteration envelope (Berry et al., 1992; Paulick et al., 2001). The stratigraphically highest deposit, Liontown, is hosted at the contact between the Trooper Creek and Rollston Range Formations; the other deposits (Handcuff, Waterloo-Agincourt, Magpie and Highway-Reward) are hosted at various stratigraphic levels within the Trooper Creek Formation (Berry et al., 1992). With one exception, all VHMS deposits in the Seventy Mile Range Group are characterised by pyritic quartz-sericite and quartz-chlorite alteration assemblages (Berry et al., 1992). The small, though very high-grade Waterloo deposit (0.372 Mt @ 19.7% Zn, 2.8% Pb, 3.38% Cu, 2.0 g/t Au and 94 g/t Ag; Hutton and Withnall, 2007) is characterised by a pyritic sericite-quartz±pyrophyllite alteration assemblage (Huston et al., 1995; Monecke et al., 2006), suggesting affinities with high sulphidation VHMS and related deposits such as at Mount Lyell in western Tasmania. All VHMS deposits in the Seventy Mile Range Group contain barite.

The Balcooma Metavolcanic Group (Fig. 52) comprises rhyolitic metavolcanic, metasedimentary rocks and minor metamorphosed mafic volcanic rocks that have been metamorphosed to lower to middle amphibolite grade (Hutton and Withnall, 2007). It is likely that metamorphosed felsic volcanic rocks that underlie the Balcooma and Dry River South deposits (Huston et al., 1992) correlate with the Mount Windsor Volcanics (Hutton and Withnall, 2007). Metasedimentary rocks that overlie Dry River South and host the Balcooma deposit may correlate with the basal part of the Trooper Creek Formation. Felsic to intermediate volcanic rocks and sedimentary rocks to the west (Huston et al., 1992) probably equate to the middle to upper parts of the Trooper Creek Formation. This interpretation is supported by SHRIMP U-Pb zircon analyses of felsic a volcanoclastic lens underlying the Balcooma deposit, which gave an age of ~480 Ma (M. Fanning, unpub. data, in Rae, 2000) and of unaltered quartz-feldspar porphyry sills, which intruded the Balcooma deposit and gave an age of ~472 Ma (Withnall et al., 1991).

The Balcooma Metavolcanic Group contains two moderate-sized VHMS deposits, the Balcooma and Dry River South-Surveyor deposits, which collectively contain 6.22 Mt grading 5.70% Zn, 2.48% Pb, 1.98% Cu, 0.63 g/t Au and 57 g/t Ag (Hutton and Withnall, 2007) in addition to a low grade massive pyrite body

at Boyds. These deposits are localised along the transition between underlying metavolcanic rocks and overlying metasedimentary rocks near the base of the sequence. The Balcooma deposit is associated mainly with chloritic alteration assemblages and lesser quartz-muscovite assemblages, whereas the Dry River South-Surveyor deposit is associated with pyritic quartz-muscovite assemblages and relatively minor chlorite-rich assemblages (Huston et al., 1992). Unlike deposits in the Seventy Mile Range Group, the Balcooma and Dry River South-Surveyor deposits lack barite. However, a small baritic prospect, West Boyds Creek, is present in the younger rocks to the west.

The Eland Metavolcanics (Fig. 52) include a sequence of andesitic to basaltic volcanoclastic rocks with minor marble and chert, which have been metamorphosed to upper greenschist grade and are possibly correlated with the Trooper Creek Formation (Hutton and Withnall, 2007). Although no significant prospects are known in this belt, correlation with the Trooper Creek Formation suggests significant potential for VHMS deposits exist.

Stolz (1995) interpreted the Seventy Mile Range Group to have formed in an evolving back-arc basin developed by extension of continental lithosphere. The earliest mafic dykes in the Puddler Creek Formation represent an alkaline intraplate association formed during the extension of a continental margin as subduction initiated. Emplacement of the Mount Windsor Volcanics followed as extension continued and a back-arc rift developed. Neodymium isotope data suggests that the felsic volcanic rocks of this unit formed in part by melting of Precambrian crust (Stolz, 1995). Continued extension eventually resulted in mantle melting to produce the felsic to intermediate volcanic rocks of the Trooper Creek Formation. The sedimentary rocks of the Rolston Range Formation represent reworking of volcanoclastic material from the Trooper Creek Formation (Stolz, 1995).

3.2.2. Porphyry and epithermal Cu-Au deposits, Macquarie Arc

Volcanic belts of Middle to Late Ordovician age in the Macquarie Arc contain a number of moderate size porphyry Cu-Au districts as well as both high sulphidation and low sulphidation gold deposits (Fig. 53). The largest of the porphyry Cu-Au districts is the Cadia district, which has global resources of 4.24 Mt Cu and 980 t Au, mostly hosted in porphyry deposits, but also in associated magnetite skarns (Cooke et al., 2007). In addition to this district, the Northparkes district has global resources of 1.57 Mt Cu and 70 t Au. These districts are hosted by ~440 Ma shoshonitic volcanic centres in which they are associated with alkalic monzonitic intrusive complexes (Crawford et al., 2007b). In the Northparkes district mineralisation was associated with the emplacement of six intrusive phases (Lickford et al., 2003), with the earliest and latest phases either weakly mineralised or barren. The mineralised pipes at Northparkes are only 100-200 m across though they extend vertically for up to 1 km (Cooke et al., 2007). Both the intrusive pipes and the country volcanic rocks are mineralised.

In the Cadia district, Cu-Au mineralisation is associated with composite intrusive pipes and dyke swarms (Wilson et al., 2003). Although much of the ore is hosted in the composite intrusive bodies, significant mineralisation in the Cadia district extends into the country rocks, mineralising both intermediate volcanoclastic rocks of the Forest Reef Volcanics and limestone to form stockwork zones and magnetite skarns. The Big Cadia deposit and other lodes mined early last century are examples of these skarns.

In both districts, Cu and Au are hosted by quartz-sulphide±magnetite±carbonate veins and breccias, with the principal ore minerals being bornite, chalcopyrite and gold. These veins occur both as stockworks (e.g., Northparkes and Ridgeway) and as sheeted vein sets (e.g., Cadia Hill and Cadia East; Cooke et al., 2007). With the exception of the Cadia Hill deposit, individual deposits have bornite-rich cores that grade outwards to chalcopyrite-rich zones and outer pyritic halos (House, 1994; Wilson et al., 2003). In these deposits Au is most closely associated with bornite. However, in the Cadia Hill deposit, the zonation is characterised by an upper bornite-rich zone that grades downward into a chalcopyrite-rich zone and then into a pyrite-rich zone. In this case, Au correlates with chalcopyrite (Holliday et al., 2002).

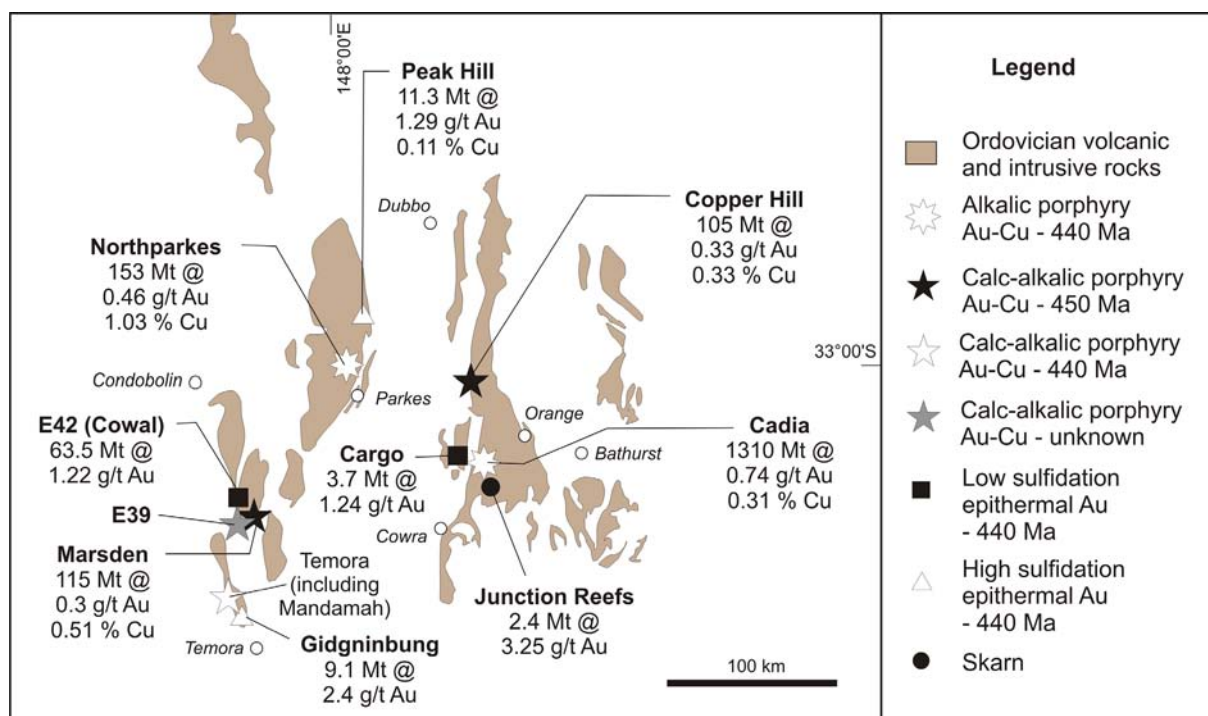


Figure 53. Geology of central New South Wales showing the distribution of Ordovician volcanic belts of the Macquarie Arc and important porphyry Cu-Au and related deposits (modified after Cooke et al. 2007).

As noted by Cooke et al. (2007), alteration associated with alkalic porphyry Cu-Au deposits differs from that typically developed around calc-alkalic porphyry Cu deposits (e.g., Lowell and Guilbert, 1970) in that high level phyllic and advanced argillic assemblages are weakly or not developed. Rather, these deposits are dominated by proximal Na-K-Ca alteration assemblages. In both the Cadia and Northparkes districts, the earliest formed assemblages are albite, biotite-magnetite and actinolite-biotite-orthoclase-magnetite assemblages (Wilson et al., 2003), typically in this order. The ores are commonly associated with the orthoclase-bearing assemblage, which formed in the selvages of mineralised veins (Heithersay and Walshe, 1995; Wilson et al., 2003). In addition, Wolfe (1994) documented a late-stage sericite-carbonate-albite assemblage at the E48 deposit in the Northparkes district, which is associated with high-grade Cu-Au-As. Importantly, unlike calc-alkalic deposits, the extent of these proximal assemblages is limited, making proximal alteration zoning less useful as an exploration vector.

Like calc-alkalic systems, the distal alteration assemblages in alkalic porphyry Cu-Au systems are marked by extensive development of propylitic assemblages. At Northparkes, this includes epidote-chlorite-carbonate-hematite-fluorite-pyrite±chalcopyrite assemblages (Cooke et al., 2007). In the Cadia district, propylitic assemblages have a complicated distribution. At the Cadia Hill deposit, epidote locally occurs in quartz-chalcopyrite±bornite±chalcocite veins. In addition to this inner propylitic zone, a regional propylitic zone surrounds deposits in the Cadia district (Tedder et al., 2001). In the upper levels of the Cadia East deposit, disseminated volcanic-hosted Cu-Au deposits are overprinted by pervasive sericite-albite-orthoclase-pyrite-tourmaline assemblages, and albite-pyrite-quartz assemblages are present above and peripheral to several other deposits (Wilson, 2003; Cooke et al., 2007).

The only other known deposit associated with alkalic magmatism in the Macquarie Arc is the Peak Hill deposit (11.3 Mt @ 0.11% Cu and 1.29 g/t Au), located to the north of the Northparkes district (Fig. 53). This deposit is characterised by pyrophyllite- and alunite-rich alteration assemblages that are quartz-deficient, leading Masterman et al. (2002) to suggest this deposit formed the roots of a high sulphidation epithermal system, with the bulk of the deposit eroded. Although Perkins et al. (1995) obtained $^{40}\text{Ar}/^{39}\text{Ar}$ ages of ~410 Ma for alteration assemblages associated with this deposit, they interpreted these ages to have been reset by post-ore deformation, preferring an age similar to the that of the Northparkes district and Lake Cowal deposits (~440 Ma).

In addition to the ~440 Ma alkalic deposits, the Macquarie Arc contains several smaller, calc-alkalic deposits such as Mandamah, Cargo, Copper Hill and Marsden (Fig. 53). Of these, the Marsden and Copper Hill resources (115 Mt @ 0.51% Cu and 0.30 g/t Au [Cooke et al., 2007] and 133 Mt @ 0.32% Cu and 0.28 g/t Au [http://www.goldencross.com.au; accessed 5 July 2008], respectively) are the most significant, although still small relative to the Cadia or Northparkes systems. In comparison with the alkalic deposits, the calc-alkalic deposits are poorly described, with the best description that of the Copper Hill deposit. Intrusive phases associated with the Copper Hill and Marsden deposits have been dated at ~450–447 Ma and ~447 Ma, respectively (Perkins et al., 1995; Cooke et al., 2007; Crawford et al., 2007b), although the errors associated with the Marsden date allow this deposit to be coeval with the ~440 Ma alkalic deposits. Hence, at least one calc-alkalic deposit, Copper Hill, is somewhat older than the alkalic deposits. However, porphyry style Cu-Au deposits in the Gidginbung Volcanics, including the Mandamah deposit, most likely has an age similar to the nearby Gidginbung (Temora) high-sulphidation Au-Cu deposit at ~436 Ma (Lawrie et al., 2007), and the E39 prospect near the Lake Cowal deposit (see below) has an apparent age ~440 Ma (Perkins et al., 1995), suggesting that calc-alkalic porphyry Cu-Au mineralisation also overlapped in time with the alkalic systems. However, known ~440 Ma calc-alkalic porphyry and related epithermal deposits are restricted to volcanic belts in the Cowal-Temora region, south and west of the belts hosting the alkalic systems.

At the Copper Hill deposit, Girvan (1992) recognised five alteration stages, an early stage quartz-magnetite-chlorite assemblage, followed by a quartz-sericite-pyrite assemblage, a sericite-chlorite-calcite-clay assemblage, followed by overprinting argillic and propylitic assemblages. High-grade quartz-pyrite-chalcopryrite veins are associated with the sericite-chlorite-calcite-clay assemblage, although some chalcopryrite and molybdenite are associated with the early quartz-magnetite-chlorite assemblage and chalcopryrite, bornite and chalcocite are associated with quartz-kaolinite veins that form part of the overprinting argillic assemblage. Torrey and Burrell (2006) suggested the association of diginite and hypogene chalcocite in these veins implies an advanced argillic overprint to this system.

The ~436 Ma (Perkins et al., 1990), calc-alkalic Gidginbung Volcanics, which are located to the north of Temora (Fig. 53), host the Gidginbung high sulphidation epithermal Au deposit in addition to porphyry Cu-Au prospects including the Mandamah and other prospects (Mowat, 2007). The Gidginbung deposit (9.1 Mt @ 2.4 g/t Au) is characterised by advanced argillic alteration assemblages. Thompson et al. (1986) described primary ore as hosted by fractured and brecciated, silicified andesitic rock that contains stringer and disseminated pyrite. This silicified zone also contains chalcedonic and cockscomb-textured quartz veins as well as minor pyrophyllite, diaspore, alunite and barite. This zone is surrounded by an advanced argillic zone containing pyrophyllite, quartz and alunite, with lesser diaspore, pyrite and kaolinite (Thompson et al., 1986).

The calc-alkalic porphyry Cu-Au systems in the Gidginbung Volcanics are characterised by classic alteration zonation: an inner hematite-magnetite-chlorite-albite-sericite-K-feldspar±biotite core grades into a Cu-Au mineralised magnetite-chlorite-albite-sericite±actinolite intermediate zone and then into outer albite-sericite-chlorite and sericite-chlorite-epidote assemblages (Mowat, 2007). Copper is hosted mostly by early, high temperature quartz-magnetite±K-feldspar±pyrite±chalcopryrite veins, although it is also present in quartz-carbonate-chlorite±chalcopryrite veins that Mowat (2007) interpreted to be remobilised. SHRIMP U-Pb dating of dykes that cut both the Gidginbung and Mandamah deposits yielded ages of ~436 Ma, consistent with zircons from Gidginbung (Lawrie et al., 2007) and an earlier SHRIMP date from a subvolcanic intrusion to the Gidginbung Volcanics (Perkins et al., 1995). $^{40}\text{Ar}/^{39}\text{Ar}$ ages of between 417 and 401 Ma reported by Perkins et al. (1995) were interpreted as resetting by post-ore deformation. Goldminco have reported a total resource (NI 43-101 compliant) of 142 Mt grading 0.33% Cu and 0.29 g/t Au for six prospects (including Mandamah) in the Gidginbung Volcanics, of which the Yiddah prospect is largest (61.2 Mt @ 0.35% Cu and 0.13 g/t Au: www.goldminco.com [accessed 23 January 2009]).

The Lake Cowal (E42) epithermal Au deposit (63.5 Mt @ 1.22 g/t Au) is located further to the north along the western margin of the Macquarie Arc (Fig. 53), close to the E39 porphyry system (Cooke et al., 2007).

This deposit, which has an age of ~439 Ma (Perkins et al., 1995), consists of auriferous quartz-carbonate-pyrite-sphalerite-galena veins associated with sericite-carbonate-pyrite alteration assemblages. Cooke et al. (2007) classified this deposit as an example of either the pyrite-quartz or carbonate-base-metal subdivisions of epithermal gold deposits, as defined by Corbett and Leach (1998), and suggested that it formed at greater depths than typical low sulphidation epithermal deposits.

3.2.3. Mineralisation and the magmatic evolution of the Macquarie Arc

Studies by Percival and Glen (2007) and Crawford et al. (2007b) have identified four chronologically and compositionally distinct phases in the Macquarie Arc. Of these the first two are not known to be associated with significant mineralisation, whereas the last two contain significant deposits.

The alkalic porphyry Cu-Au and related deposits, which constitute the major Cu-Au resource in New South Wales, are associated with the fourth, and youngest, magmatic phase of the Macquarie Arc, which has an age range from 457 to 438 Ma, with the bulk of activity in the younger part of this age range (Percival and Glen, 2007). This magmatic phase is dominated by relatively evolved, shoshonitic lavas and slightly younger porphyritic intrusions (Blevin, 2002; Crawford et al., 2007b). In most cases the alkalic porphyry Cu-Au deposits are associated with these younger intrusions.

Most calc-alkalic porphyry Cu-Au and related deposits are associated with the third magmatic phase of the Macquarie Arc, which Crawford et al. (2007) ascribed calc-alkalic affinities. Most of these magmatic rocks were emplaced between 451 and 448 Ma (Percival and Glen, 2007), although Crawford et al. (2007b) identified a dacite from the western Junee-Narromine Volcanic Belt as having an age of ~443 Ma. Although most calc-alkalic porphyry Cu-Au deposits are associated with this older period of magmatism, data in the Temora area suggests that the calc-alkalic porphyry Cu-Au and associated high sulphidation Au deposits have an age of ~440 Ma (Mowat, 2007). This suggests that a younger, ~443-440 Ma, belt of calc-alkalic magmatism and associated mineralisation may be present along the western margin of the Macquarie Arc, spatially and possibly temporally separate from other calc-alkalic systems in the Molong and Rockley-Gulgong Volcanic Belts to the east.

3.2.4. Lode Au deposits, Victorian goldfields (455-435 Ma deposits)

Until the latter half of last century, the Victorian goldfields were the most prolific Au producers in Australia (being overtaken by the Eastern Goldfields province relatively recently). The Victorian goldfields have produced nearly 2500 t of Au from discovery in 1851 to 1997 (Phillips and Hughes, 1998), with most from the Bendigo Zone (Fig. 54), and lesser production from the Stawell and Melbourne Zones. $^{40}\text{Ar}/^{39}\text{Ar}$ studies of alteration minerals (Foster et al., 1998; Bierlein et al., 2001a; Arne et al., 2001) associated with auriferous veins suggest a prolonged period of mineralisation, with a total range of 455 to 365 Ma. However, mineralisation appears to also have been episodic with the most significant event at 455-435 Ma (mostly ~440 Ma), and lesser events at 420-410 Ma and 380-365 Ma (Phillips et al., 2003). This section describes the earliest, and largest, Au event, which includes most production from the Bendigo, Ballarat and Stawell goldfields.

The ~440 Ma lode Au event is best developed in the Bendigo Zone (Fig. 54). Total Au production in the Bendigo (Ballarat) Zone was 2000 t (Phillips and Hughes, 1998), 80% of the total for the Victorian goldfields. Of this total production, 1278 t were derived from the Bendigo, Ballarat and Castlemaine goldfields, with the Bendigo goldfield accounting for nearly 700 t. In the Bendigo goldfield, auriferous veins are associated with anticlinal closures where they form saddle reefs in fold closures within Early Ordovician turbidites of the Castlemaine Group (Cherry and Wilkinson, 1994). In addition, bedding-parallel veins, transgressive veins in reverse faults, tensional veins in fold closures, and veins parallel and perpendicular to fold axes also contain Au (Turnbull and McDermot, 1998). These veins are typically quartz dominated with variable ankerite, sericite and chlorite. Ore-related sulphide minerals include arsenopyrite, pyrite, sphalerite and galena, with minor pyrrhotite, chalcopyrite and bournonite. These veins

are associated with a phyllic (sericitic) alteration halo (Turnbull and McDermot, 1998). The auriferous veins are hosted mainly in sandstone units.

Gold deposits of the Ballarat goldfield are also hosted by the Lower Ordovician Castlemaine Group (Taylor, 1998), however, in this case auriferous quartz veins are hosted by moderate-dipping faults zones with minor displacement that transect early, upright folds. The lodes are hosted with zones of dilatancy within the fault zones and consist of quartz veins with nuggety gold associated with pyrite, arsenopyrite, galena, sphalerite, pyrrhotite, chalcopyrite, stibnite and marcasite. The veins are surrounded by silica-carbonate-chlorite-sericite alteration zones (Taylor, 1998).

In contrast, Au deposits in the Stawell goldfield are hosted by metabasalt, volcanogenic sedimentary rocks, and psammopelitic rocks (Watchorn and Wilson, 1989). Structural and geochronological studies of this goldfield indicate the presence of two contrasting styles of mineralisation, which apparently formed as two temporally discrete events. At the Magdala deposit, an example of the earlier event, gold is present in southwest-dipping lodes characterised by laminated and massive quartz veins with minor pyrrhotite, pyrite, galena, sphalerite, chalcopyrite, magnetite, chalcocite, bornite, sphalerite, galena, tennantite and gold (Watchorn and Wilson, 1989; Wilson et al., 1999). Pyrrhotite and gold are intergrown with chlorite and stilpnomelane in the veins (Watchorn and Wilson, 1989). The minimum age of mineralisation is constrained by a post-ore felsic dyke, which has an age of 413 ± 3 Ma (Arne et al., 1998). Direct ^{40}Ar - ^{39}Ar dating of ore-related sericite yielded ages of ~ 438 Ma (Foster et al., 1998; D Foster, unpub. data, in Phillips et al., 2003). The characteristics of the later Au event, as represented by the Wonga deposit, are described in [section 3.3.9](#).

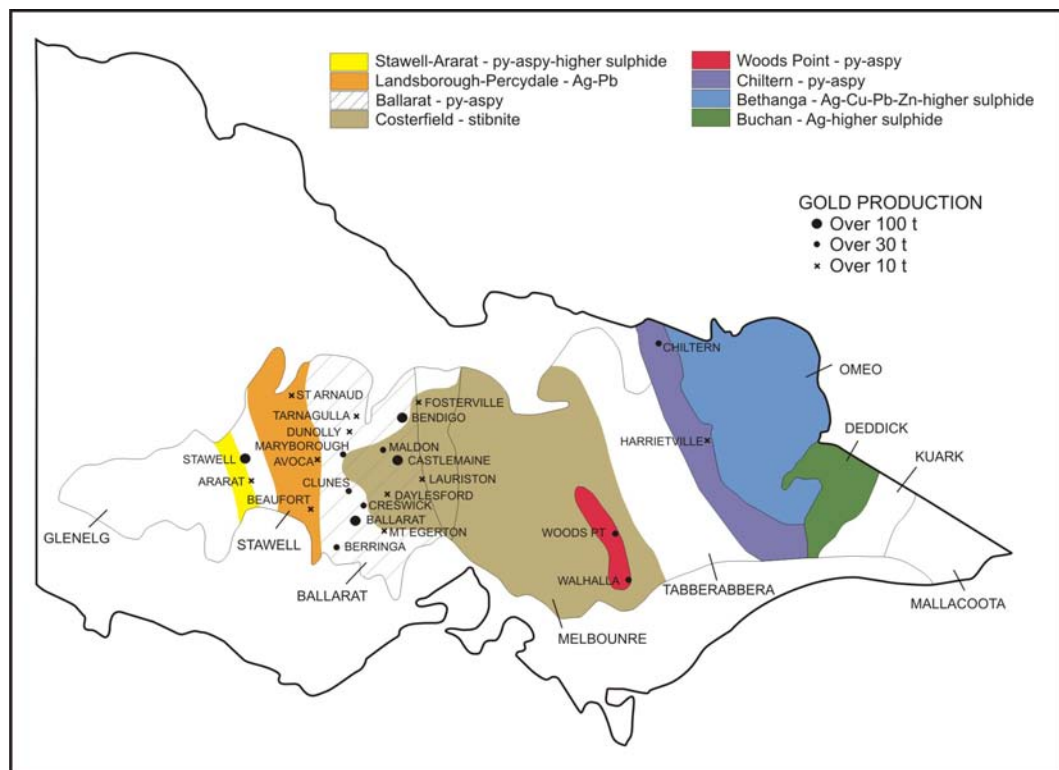


Figure 54. Geology of Victoria showing structural zones and lode gold deposits of the Victorian Goldfields (modified after Phillips et al., 2003).

Direct age constraints on the age of mineralisation in the Bendigo, Ballarat, Stawell and other goldfields are provided by Foster et al. (1998), Bierlein et al. (2001a), and Arne et al. (2001). These workers demonstrated, using mostly ^{40}Ar / ^{39}Ar analyses of hydrothermal minerals (mainly sericite and muscovite), that these deposits formed at ~ 440 Ma (total range: 455–435 Ma) in association with the Benambran Orogeny. Field relationships suggest that the deposits formed during or just after this deformation event.

Deep crustal seismic data, collected in 2006 across the southern Lachlan Orogen (Willman et al., in prep.) suggest crustal-scale controls on lode Au mineralisation in the Victorian goldfields. Based on these data, the Stawell and Bendigo Zones are the upper parts of a “V-shaped” crustal domain bounded by the moderately east-dipping Moyston Fault to the west and the moderately west-dipping Mount William Fault Zone to the east (Fig. 55). This V-shaped domain extends to the Moho and is filled by Cambrian mafic volcanic rocks at the base and Cambro-Ordovician turbidities at the top (Cayley et al., in prep.). Both to the east and west, this domain is bounded by Proterozoic rocks, at least in the middle and lower crust (Fig. 55). Based on these relationships seen in seismic data and exposures of the inferred Cambrian mafic volcanic rocks along first-order faults, the Stawell and Bendigo Zones are interpreted as thickened oceanic crust (Cayley et al., in prep.; Willman et al., in prep.). A possible interpretation is that these rocks formed in an extensional basin that was inverted and thickened during the Benambran Orogeny and later deformation events. The original extensional faults would have been reactivated during these events as reverse faults.

Spatially, most gold in the Stawell and Bendigo Zones is broadly associated with first order faults, although these faults are generally unmineralised (Willman et al., in prep.). The most significant goldfields in the Stawell Zone are in the hanging wall of the Moyston Fault and most gold in the Bendigo Zone is associated with west-dipping back thrusts (Fig. 55). In detail the Au deposits are associated with second-order faults and associated folds. Willman et al. (in prep.) interpret that the first order faults acted as fluid conduits that accessed mafic volcanics deeper in the V-shaped crustal domain, with the second order faults focussing fluid flow into favourable structural and/or stratigraphic traps. The mafic volcanic and their interflow sedimentary rocks are Au-enriched and have been interpreted as source rocks for the lode Au deposits (Hamlyn et al., 1984; Bierlein et al., 1998).

3.2.5. Geodynamic environment, and temporal and spatial zonation of 450-435 Ma mineral deposits

The period between 450 and 435 Ma is the most richly mineralised periods in the evolution of Australia. As discussed by Squire and Miller (2003), during this period the Lachlan Orogen was characterised by two quite distinct groups of mineral deposits that temporally overlap but are spatially and tectonically distinct. The northern group, which is hosted by the Macquarie Arc in the Eastern Lachlan, is characterised by porphyry Cu-Au deposits and related epithermal deposits. The southern group, which is mostly located in the Western Lachlan but may extend northward towards the southern margin of the Thomson Orogen (see section 3.2.7), is characterised by lode Au deposits. In both groups mineralisation extended from ~450 to ~435 Ma, with the main pulse at ~440 Ma; for all practical purposes the two groups are coeval.

Within the porphyry-related group, there appear to be temporal and additional spatial variations in the associated magmatism. Geochronological data suggest two discrete porphyry Cu-Au events: an older, though minor, calc-alkalic-related event at ~450 Ma (Copper Hill and probably Marsden) and a more significant, though younger, event at 440-435 Ma, which includes the vast majority of known porphyry Cu-Au resources in New South Wales. This latter event, though dominated by alkalic systems, appears to include calc-alkalic systems in the southern volcanic belt in the west and the alkalic systems developed in the central and eastern parts of the Macquarie Arc (Fig. 53). The lode Au group appears to be more homogeneous, without apparent systematic differences in deposit type or age; most deposits appear to have formed at 440-435 Ma.

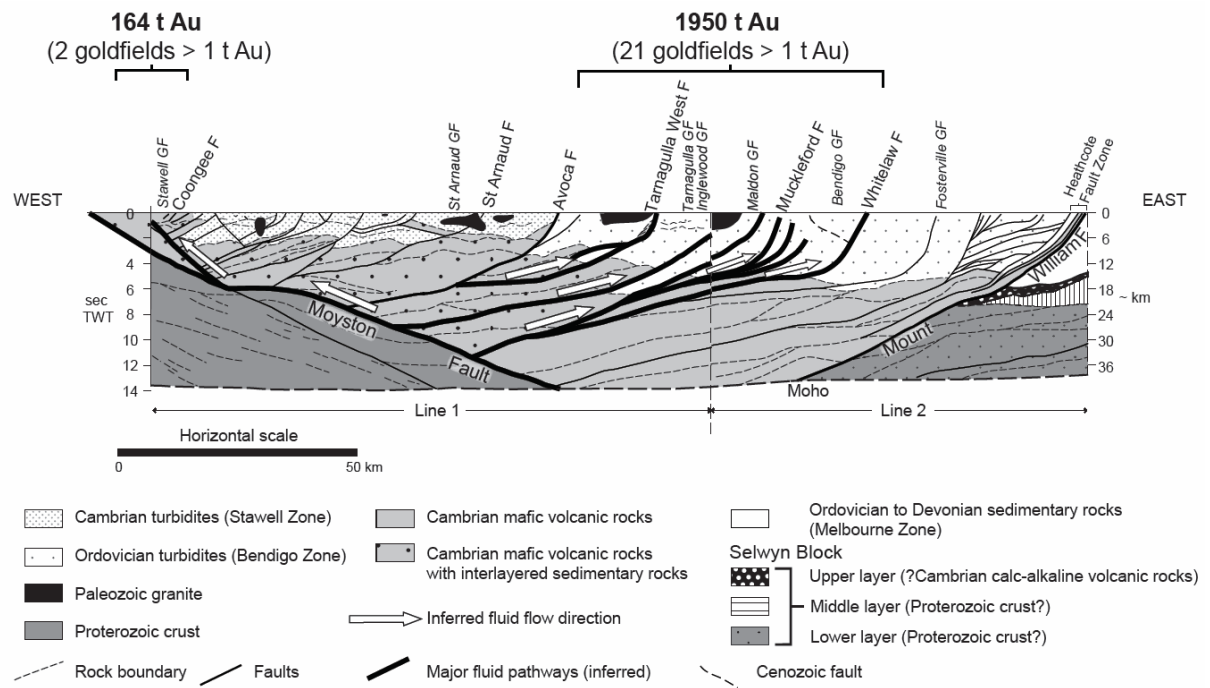


Figure 55. Interpreted cross sections along seismic lines 1, 2 and 3 showing first order faults in distribution of major lithological units (after Willman et al., in prep.).

Porphyry Cu deposits are generally associated with volcanic arcs, with Au-rich deposits commonly forming in island arcs. This environment is consistent with the geochemical and isotopic characteristics of the Macquarie Arc (Cooke et al., 2007; Crawford et al., 2007b), although in detail the alkalic porphyry Cu-Au systems formed during post-accretionary extension of this arc (Glen et al., 2007a). In contrast, most lode Au deposits form in orogenic belts, commonly late during major orogenic events (Goldfarb et al., 2005). Again, structural timing relationships for Benambran Au in the Victorian goldfields are consistent with this timing (Willman et al., in prep., and references therein), resulting in the apparent conundrum of two contrasting though contemporaneous deposit types forming in different tectonic environments.

Squire and Miller (2003) proposed that this apparent conundrum is the result of the attempted subduction of a seamount or micro-continental block by the Macquarie Arc. In this model, a microcontinent or a seamount collided with the southern part of the Macquarie Arc at ~455 Ma, resulting eventually in the lock-up of the arc at about 440 Ma. In the northern part of the arc, the Eastern Lachlan, rollback and associated slab tearing associated with this shut-down resulted in back-arc extension, and melting associated with asthenospheric upwelling produced alkalic magmatism and associated porphyry Cu-Au deposits. In contrast, collision with the microcontinent/seamount resulted in compression inboard of the arc, deforming the Bendigo and Stawell Zones of the Western Lachlan and forming lode Au deposits. This model is only valid if the western, Central and Eastern Lachlan were geographically close during the Benambran Orogeny. Most reconstructions of the Lachlan Orogen, however, suggest that the Western Lachlan was separate from Central and Eastern Lachlan during the Benambran Orogeny (e.g., Vandenberg et al., 2000; Glen, 2004; Gray and Foster, 1997, 2004: [section 2](#)), only coming together in the Tabberabberan Orogeny.

Other tectonic models have been proposed for the Lachlan Orogen during the Benambran and Tabberabberan cycles ([section 2](#)), and explain amalgamation of the Western, Central and Eastern Lachlan. For example, Powell (1984) and more recently Vandenberg et al. (2000) and Willman et al. (2002) have proposed that a major dextral strike-fault, the Baragwanath Transform (Vandenberg et al., 2000; Willman et al., 2002), separates the Western Lachlan (including the Selwyn Block and West Tasmania) from the Central and Eastern Lachlan. Vandenberg et al. (2000) inferred a displacement of 600 km along this transform prior to juxtaposition and amalgamation of these terranes during the Tabberabberan Orogeny.

This implies that the Eastern Lachlan and Western Lachlan were spatially separated, making linkage of the Western and Eastern Lachlan within a single arc system problematic.

As an alternative to the model of Squire and Miller (2003), Figure 56 presents a hypothesis in which activity along the Baragwanath Transform began prior to the Benambran Orogeny, allowing linked though contrasting tectonic environments in the Western Lachlan and Central-East Lachlan. This hypothesis is consistent with the scenario presented in Part 2 (Fig. 33) except for the inference of this transform. To the north of the transform, the Macquarie Arc formed as an island arc outboard of the Australian continent between 490 and 445 Ma (Glen et al., 2007a). The presence of 485–440 Ma calc-alkaline rocks in drill holes along the southern margin of the Thomson Orogen is interpreted by Watkins (2007) to indicate the presence of a second arc. This arc may have extended along the eastern margin of the Thomson Orogen to the Anakie Basin (Fig. 56a), where similar-aged calc-alkaline volcanic rocks are also present (section 2). If this arc was active at 480 Ma, the Seventy Mile Range Group may be in a back-arc position to it.

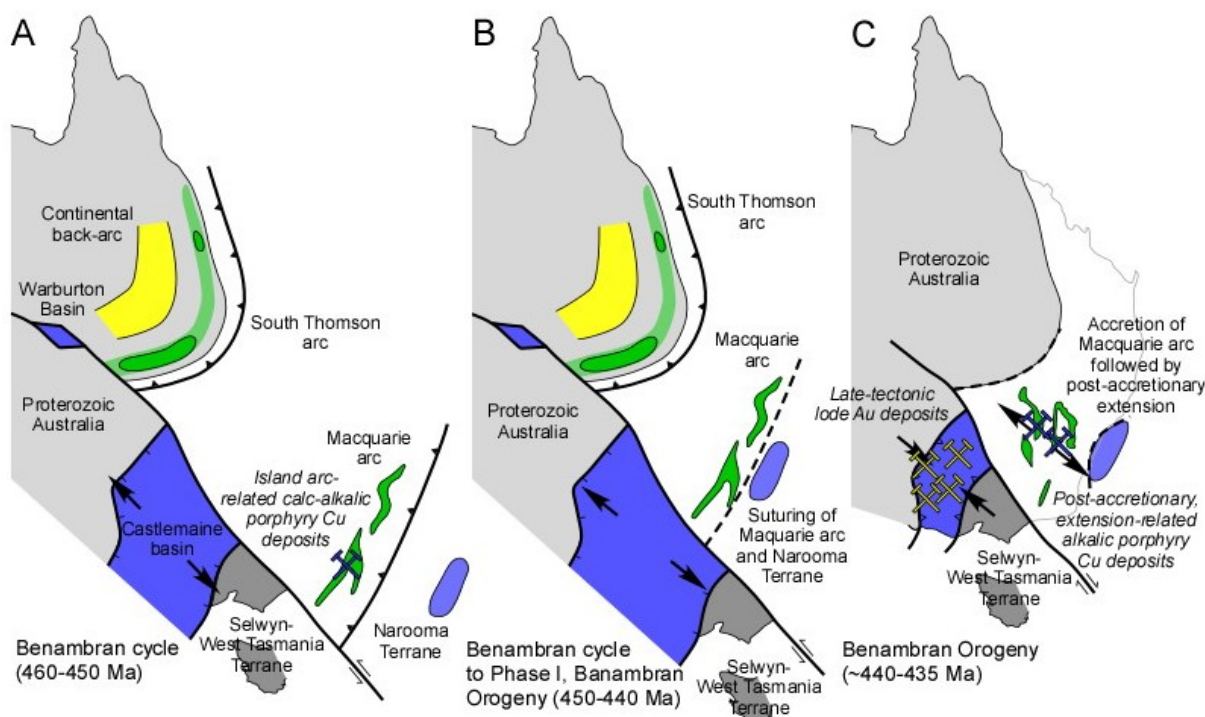


Figure 56. Schematic diagram showing possible tectonic relationship between the Macquarie Arc and the Victorian goldfields at the end of the Benambran cycle (~450–435 Ma). In this reconstruction, the Baragwanath Transform was a transfer fault between the Western Lachlan with the Central-Eastern Lachlan.

In contrast, the Western Lachlan south of the transform was characterised by ESE–WNW-directed extension resulting in the formation of a basin between the Selwyn block and the Australian continent. In the Bendigo Zone (Fig. 56a), deep-water turbidites of the Castlemaine and Sunbury Groups (“Castlemaine basin”) were deposited during the Lower to Middle Ordovician (490–455 Ma: Vandenberg et al., 2000). The contrast between convergence in the Eastern and Central Lachlan and extension in the Western Lachlan was accommodated by sinistral relative movement on the Baragwanath Transform. The only significant deposits formed during this time period are calc-alkaline porphyry deposits such as Copper Hill and Marsden in the Eastern Lachlan.

Glen et al. (2007a) inferred that west-dipping subduction below the Macquarie Arc was locked up at ~450 Ma or shortly thereafter by the subduction of a hypothesised seamount. Alternatively, as shown in Figure 56b, this subduction could have been locked by the accretion of the oceanic Narooma Terrane and/or the turbidite-dominated Eastern Lachlan. This event may have initiated Phase I (in New South Wales) of the Benambran Orogeny at ~445 Ma (c.f. Glen et al., 2007a). Upon locking of subduction associated with the

Macquarie arc, convergence between the Australian craton north of the transfer and outboard plate would have been taken up by an increased rate of subduction associated with the South Thomson arc (Fig. 56b) until the Macquarie Arc was accreted at ~443 Ma, completing Phase I of the Benambran Orogeny. Locking of both these subduction zones would have necessitated a major reorganisation of plate motion, both within the Eastern Lachlan and along the Baragwanath Transform.

Glen et al. (2007a) inferred a transient period of extension in the Macquarie Arc, from 440-435 Ma, during which time transient basins and alkalic Cu-Au deposits formed immediately following locking up of subduction associated with the South Thomson and Macquarie Arcs. This was the major consequence of this plate reorganisation in the Eastern Lachlan. It correlates yet contrasts with the main stage of compression and Au mineralisation in the Stawell and Bendigo Zones (Squires and Miller, 2003). Hence, after the accretion of the Macquarie Arc onto the Australian Craton, the geodynamic setting to the north of the Baragwanath Transform has gone from convergence to extension, whereas to the south the setting has gone from extension to compression (Fig. 56c), a change accommodated by a change in the sense of motion along this transform from sinistral to dextral. This change, and mineralisation, is most likely triggered by the accretion of the Macquarie Arc. If true, the Baragwanath Transform represents not only a major tectonic boundary, but also a major metallogenic boundary: it is unlikely that Benambran-aged lode Au deposits extend to the north and east of this transform, and porphyry Cu-Au deposits are unlikely to extend to the south and west.

3.2.6. Minor mineral deposits

In addition to the major VHMS, porphyry-epithermal Cu-Au and lode Au deposits that characterise the period between 490 and 440 Ma, small scale Zn-Pb deposits formed during this time. In Tasmania, the Oceana Irish-style Zn-Pb deposit (2.6 Mt @ 7.7% Pb, 2.5% Zn and 55 g/t Ag; Seymour et al., 2006) consists of stratiform semi-massive galena with associated siderite and sphalerite above a distinct carbonate breccia within the limestone of the Ordovician Gordon Group. Based on ore textures and relationships, Taylor and Mathison (1990) inferred a diagenetic Irish-style origin for this deposit, in which case the likely age is ~445 Ma for ore deposition. Another deposit that may have a similar origin is the Grieves Siding deposit (Seymour et al., 2006).

3.2.7. Mineral potential

Synthesis and analysis of geological and metallogenic data suggest that the Benambran cycle may have been characterised by four major geodynamic systems, in order of decreasing age:

1. Macquarie Arc;
2. South Thomson arc, associated back-arc, and extensions of these to the north;
3. Baragwanath transform system; and
4. Benambran Orogeny.

Knowledge and confidence of these tectonic systems is quite variable, with some quite well known (e.g., Macquarie Arc) and others poorly defined (e.g., South Thomson arc) and/or speculative (e.g., Baragwanath transform system). However, based on existing data, predictions can be made about the existence and likely extent of mineral systems for each of these tectonic systems.

3.2.7.1. Macquarie Arc

The potential of the 490-435 Ma Macquarie Arc for porphyry Cu-Au and epithermal Au-Cu deposits (Fig. 57) has been well established with the discovery of the Cadia and Northparkes districts as well as other smaller deposits. Potential may exist inboard of this arc for hybrid VHMS-high sulphidation epithermal Cu-Au deposits such as Mount Lyell or those in the Doyon-Bousquet-LaRonde district in Canada. In addition,

the Gamilaroi Terrane along the southwest margin of the New England Orogen has potential for porphyry and epithermal deposits. We do not consider the Western Lachlan to have significant porphyry Cu or epithermal potential of this age as it is south of the Baragwanath Transform.

Mafic-ultramafic complexes, such as the Fifield Complex and the Woolomin and Port Macquarie Terranes in the southern New England Orogen, have potential for orthomagmatic Ni-Cu-PGE accumulations, and the intrusion of younger granites into these complexes may produce hydrothermal Ni deposits analogous to the Avebury deposit in Tasmania (section 3.4.2).

3.2.7.2. South Thomson arc and northern extensions

As summarised by Watkins (2007), drill holes in the Bourke-Louth area intersected intermediate to mafic volcanic rocks with calc-alkaline to shoshonitic affinities and arc-like signatures. Limited age data suggest that these rocks were deposited between ~485 and 440 Ma, and these rocks have been interpreted as an arc developed along the southern margin of the Thomson Orogen. Additional data (see section 1.4.3) suggest that volcanic rocks of this age may extend along the eastern margin of the Thomson Orogen, through the Anakie Inlier and towards the Seventy Mile Range Group. Volcanic rocks of this age have also been intersected beneath cover in the central and western Thomson Orogen (Fig. 57), where they are interpreted to have formed during extension and crustal thinning (Draper, 2006), raising the possibility of a back-arc basin inboard from the South Thomson arc. These rocks may be the southern extension of the back arc that includes the Seventy Mile Range Group and the Balcooma Metamorphics. Hence, the available data allow the possibility that the South Thomson arc was a small part of a 485-440 Ma linked arc-backarc tectonic system that extended along the eastern margin of the Thomson Orogen from the Balcooma Metamorphics in the north to the Bourke-Louth area in the south (Fig. 57), with the majority of the system under cover.

This admittedly speculative tectonic system has significant mineral potential in addition to the known VHMS (e.g., Thalanga and Balcooma) and hybrid Cu-Au deposits (e.g., Waterloo) in the Seventy Mile Range Group and Balcooma Metamorphics. Potential for VHMS and hybrid Cu-Au deposits extends southwards into extension-related volcanic rocks undercover in the central Thomson Orogen (Fig. 57). These rocks also have potential for epithermal systems, depending on water depth, by analogy with the Drummond Basin (section 3.5.2) and possibly intrusion-related Sn-W and Au deposits. Other belts that have potential for VHMS deposits include the Broken River Province, particularly along the eastern side of the Palmerston Fault, and the Anakie Inlier (including the Fork Lagoons Beds).

In addition to VHMS potential in back-arc basins, the magmatic arc inferred along the eastern and southern margins of the Thomson Orogen also has potential for porphyry Cu-Au and epithermal deposits. The northern extent of this arc may include Ordovician intrusions in the vicinity of Charters Towers. The Anakie Inlier may also have potential for porphyry Cu deposits associated with the inferred arc. This potential extends along the eastern and southern margins into the Bourke-Louth area where arc volcanics have been intersected in drill core (Watkins, 2007).

Preliminary Nd-Sm data from mafic to ultramafic rocks the Gray Creek Complex in northern Queensland (Fig. 57) yielded a three-point $^{143}\text{Nd}/^{144}\text{Nd}$ - $^{147}\text{Sm}/^{144}\text{Nd}$ array with an apparent age of 466 ± 37 Ma (MSWD = 0.63; D Huston and R Maas, unpublished data). Although there are other interpretations of the geologic significance of this array (i.e. mixing array), the apparent age is consistent with geological relationships and suggests that these rocks have an age of ~470 Ma, with initial $\epsilon_{\text{Nd}} \sim +8$. These rocks have potential for orthomagmatic Ni-Cu-PGE deposits along with known supergene Ni-Co deposits (e.g., Greenvale). Moreover, the rift environment inferred for these rocks also has some potential for Cyprus- or Besshi-type VHMS deposits, and, where intruded by granites (e.g., Dido Granodiorite), these rocks have potential for hydrothermal Ni deposits.

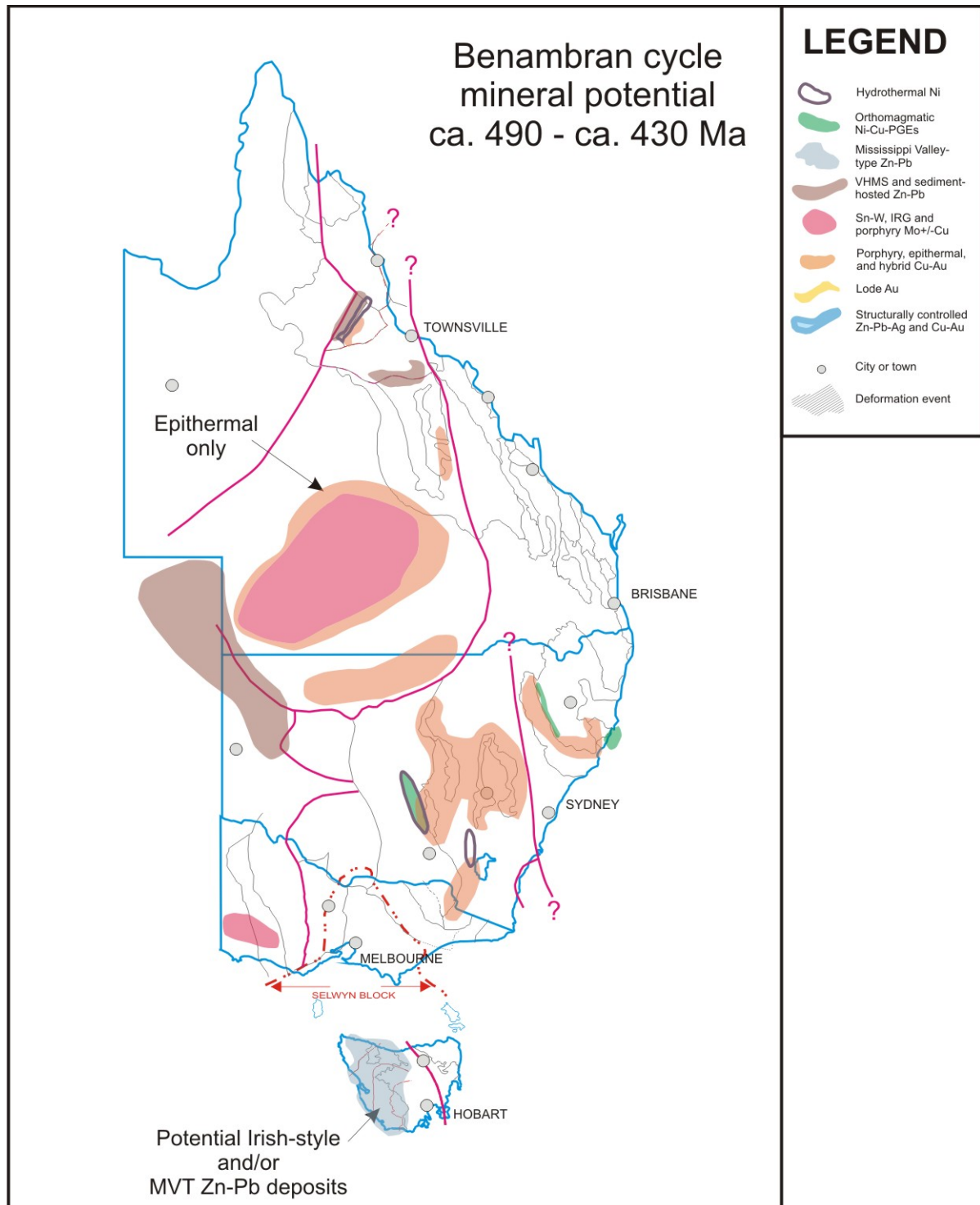


Figure 57. Mineral potential of the Benambran cycle.

3.2.7.3. Baragwanath system

As discussed above, one possible interpretation of the boundary between the Western Lachlan and Central-Eastern Lachlan is a transform. Other alternatives include a convergent margin with multiple subduction zones (Gray and Foster, 2004). If the former interpretation is correct, there would be low mineral potential associated with the Baragwanath Transform itself, although pull-apart basins associated with it (e.g., Warburton Basin) would have potential for sediment-hosted (Broken Hill-type) Zn-Pb deposits. If the Baragwanath system is a convergent margin, with ENE-directed subduction, the Eastern Lachlan would have potential for porphyry Cu and epithermal deposits just to the east of this system and possibly VHMS

deposits further inboard. However, given the uncertainty in tectonic reconstruction for this period, we consider both of these alternative mineral systems to be very speculative.

3.2.7.4. Other mineral systems

The Gordon Group in western Tasmania has known Irish-style Zn-Pb deposits, and Benambran-aged granites in western Victoria have potential for intrusion-related Sn-W and Au deposits.

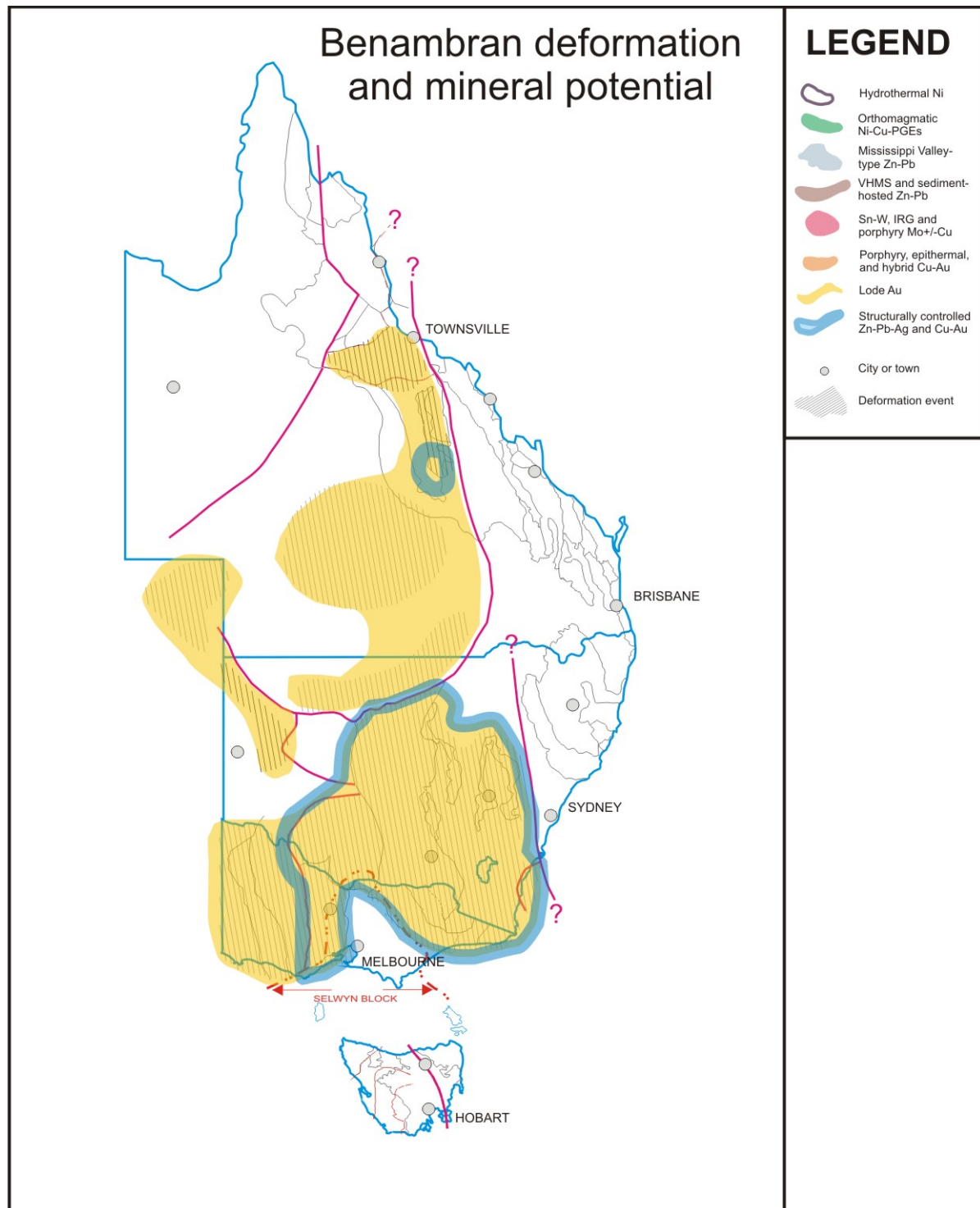


Figure 58. Mineral potential of the Benambran Orogeny.

3.2.7.5. Benambran Orogeny

Although significant known lode Au deposits associated with Benambran deformation are restricted to the Bendigo, and Stawell Zones of the Victorian goldfields, this deformation event extends from Townsville in the north to southwestern Victoria and has affected large portions of the Tasman and Thomson Orogens (Fig. 58). Despite the restricted extent of major known deposits, Benambran-aged lode Au deposits may have been more extensive. Lode Au deposits are extensively distributed across New South Wales. Although many appear to be associated with the Tabberabberan Orogeny (section 3.3.7), most are undated, raising the possibility of more extensive Benambran-aged lode Au. Gilmore et al. (2007) report that in the Koonenberry Belt, the Albert goldfield has an age of ~440 Ma (K-Ar whole rock age from altered rock associated with auriferous vein). This deposit is located on the southwest margin of the Thomson Orogen, close to the likely extension of the Baragwanath Transform. However, this transform appears to have been a major boundary between the Western Lachlan and the Central-Eastern Lachlan. Hence, although there is potential for lode Au deposits in all rocks affected by the Benambran Orogeny, we consider that the potential is much greater in the Western Lachlan, and the greatest potential for additional discoveries is under the Murray-Darling Basin to the north of the Bendigo and Stawell Zones, although this potential extends only to the Baragwanath Transform (if it actually exists). In addition, recent mapping of Ordovician rocks in the western part of the East Tasmania Terrane has identified early recumbent deformation that could be Benambran in age, raising the possibility that lode gold deposits in this area could be Benambran in age.

Potential for structurally-controlled Cu-Au and Zn-Pb deposit also exists where Benambran deformation has inverted pre-existing siliciclastic-dominated marine basins (Fig. 58). This includes much of the Lachlan Orogen and also parts of the Anakie Inlier, where Withnall et al. (1995) report an undated, structurally-controlled Cu deposit at West Copperfield.

3.3. Tabberabberan cycle (430-380 Ma)

The period between 440 and 370 Ma is characterised by a diverse assemblage of mineral deposits associated both with rifting that formed extensive basins and with compressional deformation that caused inversion of these basins. The earliest deposits in this age range are magmatic-hydrothermal deposits in the Central Lachlan in New South Wales, which range in age from 435 to 410 Ma. This age range also encompasses Middle to Late Silurian (420-410 Ma) VHMS deposits that formed in the Goulburn and Cowombat Troughs in the Eastern Lachlan and in the Hodgkinson Province in north Queensland. Following this event, lode gold deposits formed during the 420-400 Ma Bindian Orogeny. Although only weakly developed in the Victorian goldfields, this event was important in northern Queensland, when the Charters Towers goldfields formed.

Following the Bindian Orogeny, volcanism associated with the ~400 Ma Buchan Rift resulted in deposition of relatively minor syngenetic base metal and barite, epigenetic Au and porphyry-type Cu±Mo deposits in eastern Victoria. At ~385-380 Ma, hybrid VHMS-magmatic Cu-Au deposits, including the Mount Morgan deposit, formed in the Calliope Arc in the New England Orogen. During this time, epigenetic Au, Cu-Au and Zn-Pb-Ag deposits formed in response to inversion of the Cobar Basin (Lachlan Orogen). Following or possibly overlapping these events, compression associated with the Tabberabberan Orogeny was accompanied by lode gold mineralisation. This event was the major Au event in Tasmania, but only a subordinate event in the Victorian goldfields and the Hill End Trough in New South Wales.

3.3.1. Magmatic-related tin, tungsten, base metal and molybdenum deposits, Central Lachlan, central New South Wales

The Central Lachlan in central New South Wales contains a large variety of epigenetic deposits temporally and spatially associated with 435-410 Ma granites. The most significant of these are Sn deposits that form the Wagga Sn belt, which extends from northern Victoria to Kikiora in central New South Wales (Fig. 59).

However, the Central Lachlan also contains epigenetic deposits with commodities ranging from Mo (e.g., Holbrook deposit) to Cu-Zn-Pb-Ag (Tara) to W (with Sn at Kikoira and nearby deposits) to Cu-Au (Mineral Hill).

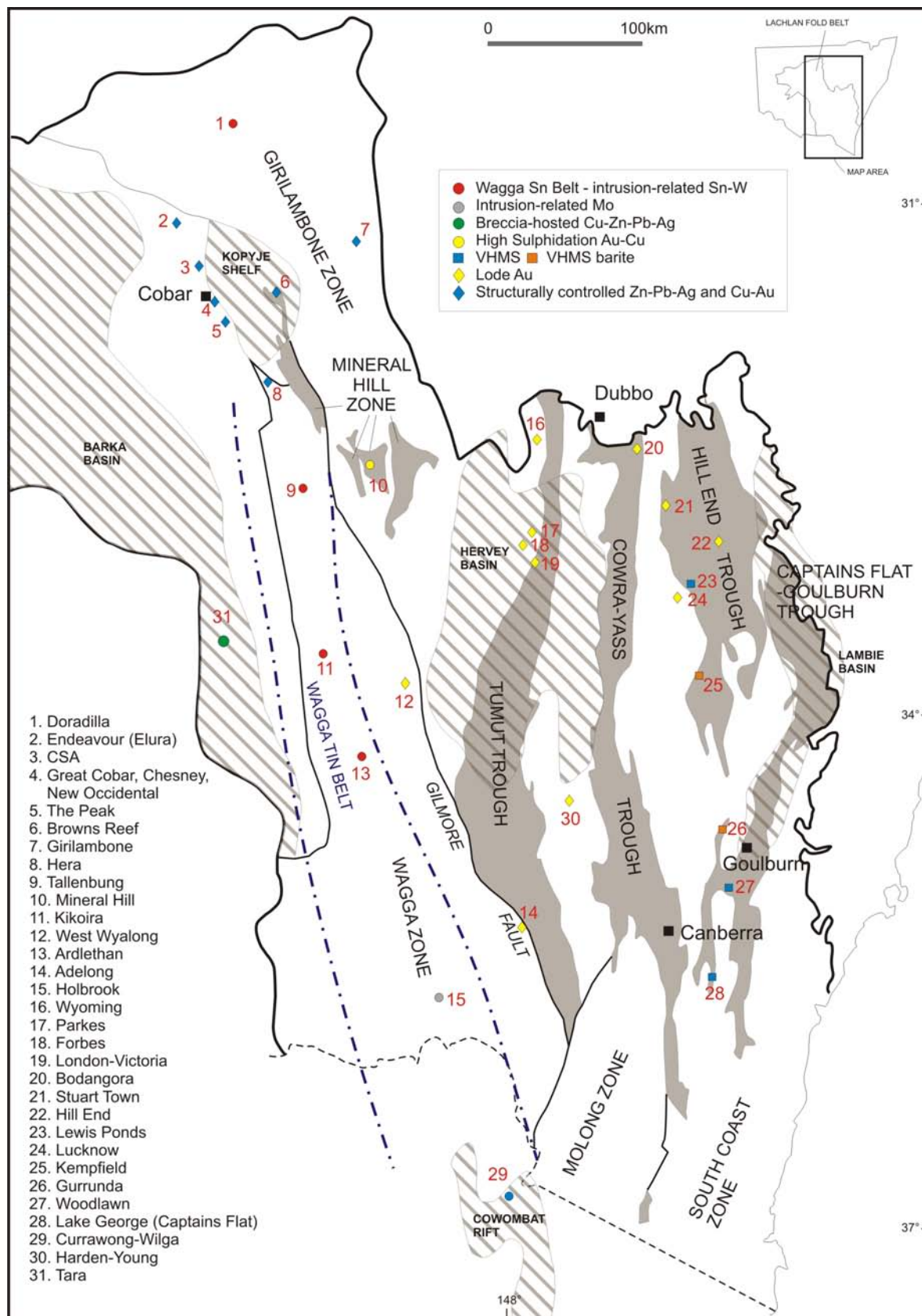


Figure 59. Distribution of selected mineral deposits in central and eastern New South Wales.

A large number of Sn prospects as well as a few small to moderate-sized Sn deposits form the Wagga Sn belt, which extends from northern Victoria to north central New South Wales (Fig. 59). These deposits are largely associated with S-type granites of the 435-410 Ma (L Black, in Colquhoun et al., 2005) Koetong Supersuite, although smaller deposits are also associated with I-type granites (Blevin and Chappell, 1995). The more significant districts in this belt include Ardlethan (31,568 t Sn production from 9 Mt of ore and a 3.025 Mt resource grading 0.42% [not JORC compliant]; Paterson, 1990), Doradilla and Tallebung (3350 t Sn-W concentrate (~2000 t Sn) from alluvials: www.ytcreources.com [accessed 7 August 2008]). However, recent geochronological studies (Burton et al., 2007) of the Doradilla district indicate that the ore-related intrusive rocks are Middle Triassic (~235 Ma) in age; hence these are discussed later (section 3.5.11).

Of the deposits in the Wagga Sn belt, the best described is the Ardlethan deposit. Ren et al. (1995) indicated that this deposit is hosted by breccia pipes within the ~417 Ma Mine granite near its contact with the strongly fractionated, ~410 Ma Ardlethan Granite (Rb-Sr ages from Richards et al., 1982). Both of these granites intrude Ordovician sedimentary rocks. The Sn is associated with tourmaline that is locally massive to semi-massive and forms part of the matrix to the breccia pipes and replaces the breccia clasts. The ores also contain local vugs filled initially by tourmaline, quartz, cassiterite and arsenopyrite, and then by clear quartz, pyrite, chalcopyrite, sphalerite and galena. In addition to tourmaline, the wall rocks to, and clasts within the breccia pipes, are altered by kaolinite-, sericite- and chlorite-bearing assemblages (Ren et al., 1995).

Deposits in the northern part of the Wagga Sn belt appear to be associated with the ~432-428 Ma Kikoiira and Erigolia Granites (age data from Black, in Colquhoun et al., 2005), which are part of the Koetong Supersuite (Colquhoun et al., 2005). These deposits consist mainly of cassiterite-bearing quartz veins that are hosted by Ordovician metasedimentary rocks of the Wagga Group or the Kikoiira Granite. Associated minerals include scheelite, wolframite, native silver, arsenopyrite, pyrite and chalcopyrite. Some veins also carry tourmaline, which is also a local component or muscovite-dominant alteration assemblages (Burton and Downes, in Colquhoun et al., 2005). Significant alluvial production has also been sourced from the Kikoiira deposits at Gibsonvale (6000t Sn, in Colquhoun et al., 2005). Further to the north a cassiterite-wolframite sheeted vein system of several hundred metres vertical extent is present at Tallebung beneath historic hard rock and alluvial operations. Mineralisation is presumably sourced from an unexposed granite at depth (P Blevin, pers. comm., 2009).

The northern part of the Wagga Sn belt also contains base metal deposits with slightly younger apparent ages. One of the better studied deposits is the Tara deposit, which was discovered by drilling an aeromagnetic anomaly. The mineralisation consists of stockwork quartz-pyrite±pyrrhotite veins and disseminated pyrite±pyrrhotite hosted by turbiditic rocks of the Wagga Group. The veins also include variable chalcopyrite, sphalerite, galena, arsenopyrite and cassiterite, with muscovite and minor chlorite and carbonate gangue (Burton and Downes, in Colquhoun et al., 2005). Alteration assemblages associated with the stockwork veins include an early tourmaline-bearing assemblage and a syn-vein silica-rich assemblage. ⁴⁰Ar-³⁹Ar data from vein-hosted muscovite yields an age of ~420 Ma, which is interpreted as the age of mineralisation (Downes and Phillips, 2006).

The most significant prospect in the northern Central Lachlan is the Browns Reef deposit, a stratabound Zn-Pb deposit hosted by the Preston Formation, which is constrained to a Pridoli to Lochkovian age (418-416 Ma; Colquhoun et al., 2005). This deposit has a JORC-compliant resource of 20 Mt grading 2.0% Zn, 1.1% Pb, 0.1% Cu and 9 g/t Ag (www.cometres.com.au, accessed 20 August 2008).

The Preston Formation comprises slate and volcanogenic sandstones with several bodies of rhyolite sills. Several facies of sandstone are present, including crystal-rich, lithic-feldspathic and quartz-lithic varieties (Colquhoun et al., 2005). The deposit is hosted by the lower part of the Preston Formation and consists of disseminations, blebs and stringers of sulphide and sulphide-bearing quartz veins and stockwork in

silicified sandstone. Sulphide minerals include pyrite, with lesser sphalerite, galena and chalcopyrite, and trace arsenopyrite, covellite and bornite. Gangue and alteration minerals associated with these veins include white mica, chlorite and carbonate (Burton and Downes, in Colquhoun et al., 2005). Burton and Downes (in Colquhoun et al., 2005) preferred an interpretation in which this deposit formed epigenetically, possibly in a similar manner to deposits in the Cobar Basin, noting similarities in Pb isotopic signatures (see [section 3.3.8](#)).

The Mineral Hill Au-Cu deposit (production to December 2003: 0.018 Mt Cu and 10.3 t Au [Morrison et al., 2004]), located to the northeast of the Tara and Kikoiira deposits, is hosted by the Mineral Hill Volcanics, which consist of rhyolitic to dacitic lapilli, crystal-lithic and vitric tuffs with minor agglomerate, lava, limestone, siltstone and sandstone (Morrison et al., 2004). This unit is overlain by the unmineralised Early Devonian (~418 Ma: Morrison et al., 2004) Talingaboolba Formation, which comprises conglomerate, sandstone and siltstone. Although undated, the Mineral Hill Volcanics may have a similar age to regional magmatic rocks, which have ages between 428 and 421 Ma (Blevin and Jones, 2004; Morrison et al., 2004).

The ore zone consists of quartz veins and breccias with an early pyrite-gold-native bismuth±arsenopyrite assemblage overprinted by a late sulphide assemblage that grades from gold-chalcopyrite-bismuthinite-bornite in the lower parts of the ore zone to sphalerite-galena-fahlore and gold-silver-arsenopyrite-stibnite in the highest part of the mine stratigraphy. The gold-chalcopyrite-bornite assemblage is associated with a magnetite-pyrite-chlorite-hematite±biotite alteration assemblage (Morrison et al., 2004).

Morrison et al. (2004) interpreted the Mineral Hill deposit to have formed at relatively high levels in the crust, as indicated by vein textures and textures within the associated magmatic rocks. They indicated that the mineralising event is constrained by the age of the oldest igneous rock in the region and the age of the Talingaboolba Formation to between 428 and 418 Ma, and inferred a magmatic affinity for mineralisation.

Near Holbrook in the very southern part of the Wagga Sn belt, Mo-bearing quartz veins are hosted by Lower Devonian granites that have been variably greisenised. In addition to molybdenite, these veins also contain cassiterite, chalcopyrite with traces of Bi and W minerals. Molybdenite, cassiterite and tourmaline are also present within greisenised zones within the host granite (Kennedy and Loudon, 1964). Re-Os dating of molybdenite from this deposit yielded an age of ~423 Ma (Norman et al., 2004).

3.3.2. VHMS and related deposits, Middle Silurian rift basins, Lachlan Orogen

After the Mount Read Volcanics and the Mesoarchaean Murchison Province, Middle Silurian rift basins in central New South Wales and eastern Victoria ([Fig. 59](#)) form the third most significant VHMS province in Australia. Collectively, production and resources from these deposits total 3.8 Mt Zn, 1.4 Mt Pb, 0.9 Mt Cu, 5.5 kt Ag and 44 t Au, with the largest Woodlawn deposit having a global resource of 23.9 Mt grading 9.6% Zn, 3.8% Pb, 1.7% Cu, 79 g/t Ag and 0.52 g/t Au (Davis, 1990; Huston et al., 1997; www.trioriginminerals.com.au [accessed 04 July 2008]).

Known deposits are hosted by Middle Silurian volcano-sedimentary rock packages within the Goulburn Basin in New South Wales and the Cowombat Rift in Victoria. Volcanic-hosted massive sulphide deposits within these rifts tend to occur in clusters of up to ten deposits. In the Goulburn Basin, many of these districts have a similar stratigraphy, as noted by Davis (1990). Turbiditic sedimentary rocks of Late Ordovician to Early Silurian age are overlain unconformably by siltstone and shale with limestone lenses of Middle Silurian age. These rocks are overlain by felsic volcanic rocks, typically quartz and feldspar-phyric rhyolitic and rhyodacitic volcanoclastic rocks. L Black (in Huston et al., 1997) reported an age of 417 ± 4 Ma for these volcanic rocks in the Lewis Ponds district. These felsic volcanic rocks are overlain by siliciclastic rocks, dominantly siltstone with minor sandstone and variable felsic and mafic volcanic rocks. Volcanic-hosted massive sulphide and related deposits appear to be localised along the upper contact of the

volcanic unit and in siliciclastic sedimentary rocks just below the lower contact of the volcanic unit (Davis, 1990).

The Goulburn Basin hosts three significant VHMS deposits, the Woodlawn, Lake George (aka Captains Flat) and Lewis Ponds deposits, as well as a number of smaller prospects. In addition to these sulphide-rich deposits, the Goulburn Basin also hosts a number of low sulphide, barite-rich deposits such as the Kempfield and Gurrundah prospects. These deposits typically contain a few percent disseminated sphalerite, galena, pyrite and tetrahedrite (Stevens, 1977). Despite the low sulphide content of these barite deposits, some appear to have potential as Ag resources. For instance, the Kempfield deposit has JORC-compliant resources of 3.72 Mt grading 44.7% BaSO₄, 0.70% Zn, 0.46% Pb and 94.7 g/t Ag; www.kempfieldsilver.com.au [accessed 25 July 2008]). These characteristics, as well as the recognition of actively-forming sulphate deposits on the seafloor at shallow depths (<600 m: Hannington et al., 2005), suggest that Goulburn Basin barite deposits may have been deposited at relatively shallow depths at low temperature. Although not recognised to date, these barite deposits may also have potential for significant Au grades.

The VHMS deposits of the Goulburn Basin typically contain multiple lenses; for instance the Woodlawn deposit contains three major lenses as well as a number of minor lenses. These lenses are dominated by pyrite with variably abundant sphalerite, galena and chalcopyrite and minor to trace quantities of arsenopyrite, tetrahedrite-tennantite, stannite, bismuthinite, marcasite, pyrrhotite, pyrargyrite, electrum, bournonite, jamesonite, bornite and scheelite (Edwards, 1943; Edwards and Baker, 1953; Markham, 1961; Ayres et al., 1979; Agnew, 2003). The Woodlawn, Captains Flat and, to a lesser extent, Lewis Ponds deposits are zoned, with stratigraphically lower parts of the ore zones enriched in Cu. At both the Woodlawn and Captains Flat deposits, barite is abundant (Ayres et al., 1979; Davis, 1975), however, at Lewis Ponds, the only evidence for sulphate minerals is silica pseudomorphs preserved in less recrystallised ores (Agnew, 2003).

The Goulburn Basin deposits are unusual for the relatively high abundance of talc as a proximal alteration mineral. McKay and Hazeldene (1987) and Agnew (2003) describe chlorite-talc-carbonate alteration assemblages proximal to the Woodlawn and Lewis Ponds deposits. In both cases, this assemblage grades outward to chlorite-sericite assemblages, with the relative proportion of sericite increasing outwards. At the Lewis Ponds deposit, Agnew (2003) described an unusual quartz-sericite±hyalophane assemblage in the immediate footwall to the Toms lens.

The Cowombat Trough in northeastern Victoria (Fig. 59) hosts two VHMS deposit in the Gibsons Folly Formation, a Middle Silurian unit of interleaved mudstone, basalt, dacite and rhyolite (Allen and Barr, 1990). The larger of the two deposits, Currawong, contains a JORC-compliant resource of 9.0 Mt grading 4.2% Zn, 0.8% Pb, 2.0% Cu, 39 g/t Ag and 1.2 g/t Au, whereas the smaller Wilga deposit contains a resource of 3.3 Mt grading 6.1% Zn, 0.5% Pb, 3.4% Cu, 36 g/t Ag and 0.5 g/t Au (www.jabirumetals.com.au; accessed 25 July 2008). These two deposits are located 3 km along strike of each other and consist of two major lenses (and two minor lenses) and a single lens, respectively. The Currawong deposit appears to have originally been a single lens that was chopped into multiple lenses by later faults (Bodon and Valenta, 1995).

Volcanic-hosted massive sulphide deposits in the Cowombat Trough are characterised by abundant pyrite, sphalerite, galena and chalcopyrite, with minor to trace arsenopyrite, tetrahedrite, pyrrhotite, bismuth minerals and electrum. Unlike the deposits of the Goulburn Basin, these deposits contain abundant magnetite, particularly in the stratigraphically lower parts of the deposits, and they contain very minor barite (Allen and Barr, 1990; Bodon and Valenta, 1995). However, like the deposits in the Goulburn Basin, these deposits contain talc in a stilpnomelane-carbonate±talc±chlorite assemblage that is closely associated with pyrite-magnetite-rich zones and with sphalerite-rich stringer zones (Bodon and Valenta, 1990). Elsewhere the ores are associated with chlorite±quartz assemblages with variable amounts of carbonate. Although complicated by faulting, Bodon and Valenta (1990) reported a broad ore zonation of massive pyrite → intermixed massive pyrite-pyritic Zn → pyritic Zn → Zn-Pb rich from stratigraphic base to top.

3.3.3. Base metal, gold and barite deposits, Buchan Rift

The Buchan Rift, in eastern Victoria contains a number of small occurrences and deposits with a diverse metallogeny. This rift consists of two broad units, the Snowy River Volcanics and the conformably to unconformably overlying Buchan Group (Vandenberg et al., 2000). The former consists of a diverse assemblage of subareal to submarine felsic volcanic rocks interbedded with siliciclastic rocks including sandstone, conglomerate and black shale. The Buchan Group consists of limestone and marl. Fossil evidence suggests that the upper part of this group was deposited during the late Emsian (~400 Ma: Vandenberg, 2003).

The Snowy River Volcanics contain minor, apparently stratiform accumulations of sphalerite, galena and chalcopyrite that are associated with barite. Although these prospects are not of economic interest, they are interpreted to indicate exhalative seafloor mineralisation (Vandenberg et al., 2000). In addition, the Snowy River Volcanics also contain weakly Au-Ag-mineralised epithermal veins, which are characterised by colloform and crustiform quartz, and auriferous quartz stockworks associated with disseminated pyrite, chalcopyrite, galena, arsenopyrite and barite. Barite is also present in massive veins (Vandenberg et al., 2000). The southern margin of the Buchan Rift also hosts porphyry-style Cu-Mo deposits, probably associated with Early Devonian stocks; the northern margin of the rift contains Pb-Zn-Ag veins. Although undated, Vandenberg et al. (2000) infer that all mineralisation within the Snowy River Volcanics is syn-rift in timing. The overlying Buchan Group contains a number of carbonate-hosted Zn-Pb deposits interpreted as post-Tabberabberan, epigenetic deposits (Arne et al., 1994).

3.3.4. Late Silurian VHMS and related deposits, Hodgkinson Province

The Hodgkinson Formation, which consists largely of monotonous siliciclastic arenite and mudstone (Bultitude et al., 1997), contains several small Cu-Zn deposits interpreted as Besshi-type VHMS deposits (Bain and Murray, 1998). This unit also contains minor conglomerate, chert, tholeiitic basalt and limestone. Fossils within the limestone lenses indicate an Early Devonian (Lochkovian: 416-411 Ma), or even Late Silurian to Late Devonian (Famennian: 375-360 Ma) age (Bultitude, in Bain and Murray, 1998). These rocks contain a number of small, apparently stratiform Cu-Zn deposits, the most significant of which include the Mount Molloy, Dianne and OK deposits. The largest production came from the Dianne deposit, which produced 18,000 t of Cu (Garrad and Bultitude, 1999); the OK deposit produced 7934 t Cu (and 97 kg Ag and 12.8 kg Au) from just over 88,000 t of ore (<http://www.axiom-mining.com>, accessed 11 August 2008); no production is recorded from the Mount Molloy deposit.

The Mount Molloy deposit is hosted by carbonaceous and pyritic shale, whereas the Dianne deposit is hosted by an interbedded shale-greywacke package. In contrast, the OK deposit is hosted by mafic volcanic rocks. At the Mount Molloy and OK deposits, massive sulphide zones are underlain by stockwork zones. The Dianne deposit also comprises massive sulphide, but lacks the stockwork zone. The main ore minerals in all three deposits are pyrite, chalcopyrite and sphalerite, with minor tetrahedrite-tennantite also present at the OK deposit (Gregory et al., 1980; Morrison and Beams, 1995; Garrad and Bultitude, 1999).

3.3.5. Epigenetic gold deposits, Braidwood-Majors Creek-Araluen district, New South Wales

Extensive alluvial Au deposits in the Braidwood-Majors Creek-Araluen district in southeastern New South Wales are spatially associated with the ~411 Ma (Bodorkos and Simpson, 2008) I-type Braidwood Granodiorite. These alluvial Au deposits appear to be derived from the erosion of relatively small quartz vein deposits hosted by this granodiorite, which itself intruded rhyolitic and dacitic volcanic rocks of the ~413 Ma (Bodorkos and Simpson, 2008) Long Flat Volcanics.

Total production from the alluvial goldfields in the Braidwood-Majors Creek-Araluen district totalled 38.75 t, with a further 0.53 t produced from high grade (14-122 g/t Au) vein deposits. A JORC-compliant resource of 3.58 Mt grading 2.8 g/t Au for a further 10.0 t Au has been reported for the Dargues Reef deposit (Glover, 2006). Glover (2006) recognised two styles of mineralisation in the Majors Creek area: (1) quartz-carbonate-sulphide and massive sulphide veins and breccias, and (2) disseminated sulphide zones associated with intense sericite-carbonate alteration assemblages. In addition to quartz, the veins contain rhodochrosite, chalcopyrite, sphalerite, native gold and silver, galena and tetrahedrite. The disseminated sulphide zones contained a similar mineralogy, but also minor tellurides, Bi minerals and pyrrhotite in addition to pyrite, which comprise up to 30% of the rock (Glover, 2006).

Potassium-argon dating of sericite closely associated with the Dargues Reef yielded ages of ~411 to ~400 Ma, suggesting that mineralisation was either contemporaneous with or closely followed the emplacement of Braidwood Granodiorite (McQueen and Perkins, 1995). Based on geochemical and isotopic analogies and the overlapping timing of mineralisation and granodiorite intrusion, Glover (2006) interpreted the hard rock deposits in the Majors Creek area as intrusion-related gold, probably associated with the Braidwood Granodiorite. Minor disseminated molybdenite is associated with the Braidwood Granodiorite near Araluen (P Blevin, pers comm., 2009).

3.3.6. Lode gold deposits, Victorian goldfields (420-400 Ma deposits)

Of the three Au events that have affected the Victorian goldfields, the 420-400 Ma event appears to have been the least significant, mostly overprinting goldfields in the Bendigo and Stawell Zones, including the Bendigo, Ballarat and Stawell goldfields (Phillips et al., 2003). The only goldfield in which this event appears to have been the major event is the Tarnagulla goldfield (Bierlein et al., 2001a). The Tarnagulla goldfield, which produced 22 t of Au, is located within the northern part of the Bendigo Zone and hosted by turbiditic Ordovician rocks of the Castlemaine Group. Mineralisation is inferred to have occurred during brittle-ductile reactivation of reverse faults that cut a north-trending syncline (Cuffley et al., 1998). $^{40}\text{Ar}/^{39}\text{Ar}$ dating of hydrothermal muscovite from this deposit yielded multiples ages of ~410 Ma and a single age of ~398 Ma. The latter age was obtained from discordant, late-stage veins that cut the main reef (Bierlein et al., 2001a).

Despite being located within 2 km of the Magdala deposit (see section 3.2.4), the Wonga deposit, in the Stawell goldfield, has quite a different structural style and, apparently, is significantly younger. The Wonga deposit, which has produced over 9 t of Au, comprises a series of 350°-trending lenticular lodes (shear lodes) hosted by two shear zones that are linked by a series of irregular “link” lodes that strike on average 320°. The shear lodes dip 25-50° to the east, whereas the link lodes dip 40-70° to the southeast. These lodes consist of massive and laminated quartz veins and quartz breccia with abundant wall rock fragments. However, locally the quartz lodes are absent and the Au is hosted in zones defined by disseminated arsenopyrite and pyrite. Mineralogically, the ores are simple, with abundant pyrrhotite, arsenopyrite, löllingite, pyrite, with minor stibnite, chalcopyrite, molybdenite, rutile, ilmenite and native bismuth (Wilson et al., 1999; Miller and Wilson, 2004).

This contrasts with the Magdala deposit, in which the Au is hosted by southwest- to west-dipping reverse faults and has a more complex ore mineral assemblage. The age of the Wonga deposit is constrained between ~423 Ma, the age of dykes overprinted by the ores (Wilson et al., 1999) and ~400 Ma, the age of emplacement of the Stawell Granite, which has thermally metamorphosed the Wonga ores (Phillips et al., 2003).

3.3.7. Lode gold deposits, Charters Towers goldfield

Although the 420-400 Ma Au event in the Victorian goldfields was a relatively minor event, a significant Au event of similar age occurred in north Queensland. The Charters Towers goldfield and nearby deposits produced over 6 Moz (180 t) Au between 1872 and 1918 in addition to major quantities of Ag, Pb and Cu

(Kreuzer, 2005). These deposits, which are mostly hosted by phases of the Ravenswood Batholith, comprise auriferous massive (buck), comb-textured and brecciated quartz veins that average 10% sulphide minerals. The sulphide minerals include pyrite, sphalerite and galena, with local arsenopyrite, chalcopyrite and tetrahedrite-tennantite and minor gold and gold tellurides. The veins are associated with narrow (to 0.1 m) alteration selvages characterised by a sericite-calcite-ankerite-pyrite assemblage (Peters and Golding, 1989).

$^{40}\text{Ar}/^{39}\text{Ar}$ data of Kreuzer (2005) from the Charters Towers and nearby Hadleigh Castle goldfields indicated ages of ~407 Ma for both goldfields (total range: 400–412 Ma), consistent with previous K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ data of Morrison (1988) and Perkins and Kennedy (1998). As noted by Kreuzer (2005), these ages overlap those of some regional granites (e.g., Deane, Carse-O-Gowrie, Chippendale and Broughton River granodiorites), although not the host Millchester Creek tonalite. This relationship, combined with Nd and Pb isotope data and granite compositional data, led Kreuzer (2005) to conclude that the Charters Towers deposits were lode (orogenic) Au rather than intrusion-related Au deposits.

Although small in comparison (20 t total production), vein Au deposits of the Etheridge goldfield (Bain et al., 1998) in the Georgetown Inlier to the northwest share many similarities with the Charters Towers goldfield. Unlike the Charters Towers deposits, the Etheridge deposits are hosted mostly by Mesoproterozoic schist, gneiss, metabasite and granite, with only minor Siluro-Devonian granite hosts. However, like the Charters Towers deposits, these deposits consist of sulphide-rich (pyrite, sphalerite, galena and chalcopyrite) quartz veins up to 5-m-wide and have Pb isotope model ages of 426–398 Ma. Even smaller goldfields to the northeast of the Etheridge goldfield, which are hosted by ~407 Ma granites also have similarities to the Charters Towers goldfields (Bain et al., 1998). This suggests that the ~410 Ma lode gold event was relatively widespread through north Queensland.

3.3.8. Lode gold deposits, central New South Wales

The Hill End Trough in central New South Wales (Fig. 59) was the first site of the first major Australian gold rush in 1851. Total production and reserves for the northern part of the trough exceed 95 tonnes, with major clusters of deposits at Hargraves (8.7 t), Hill End (21.5 t), Sofala-Wattle Flat (34.6 t) and Gulgong (19.0 t; Downes et al., 2008). The high grade character of some of these ores is illustrated by a recent announcement of a JORC-compliant resource at the Reward deposit of 0.124 Mt grading 19 g/t Au for 2.4 t Au (www.hillendgold.com.au; accessed 28 July 2008).

The deposits in the Hill End cluster are mostly hosted by interbedded sandstone and shale of the Late Silurian Chesleigh Formation (420–415 Ma; Pogson and Watkins, 1998), with smaller deposits also present in the younger Cookman Formation and Crudine Group. Deposits in the Hargraves cluster, ~30 km to the north, are hosted by the Cunningham Formation, which is dominated by slate with minor feldspathic sandstone beds (Windh, 1995). In both these clusters, the Au is closely associated with the regional, north-trending Hill End Anticline, with a few small deposits associated with adjacent anticlines. The Hill End Anticline structure is cut and Au veins are metamorphosed by the ~320 Ma (Pogson and Watkins, 1998) Bruinbun Granite (Windh, 1995), providing a minimum age constraint on mineralisation. The host units to these deposits range in age from the late Ludlow (~420 Ma) through the late Emsian (~400 Ma; Pogson and Watkins, 1998), which provides a maximum constraint on the age of mineralisation.

In detail, Windh (1995) noted that Au mineralisation is associated with bedding-parallel quartz veins that thicken into anticlinal hinges, forming saddle reefs. The Au-rich portions of these veins tend to be localised along the east-dipping limb of anticlines, with the saddle reefs generally being Au-poor. These veins consist mostly of laminated to massive quartz with subordinate calcite and chlorite and minor to trace muscovite, sulphides (pyrite, chalcopyrite, galena, sphalerite, pyrrhotite and arsenopyrite), and gold. The vein laminations contain chlorite and muscovite. Narrow leader veins, sub-horizontal, discordant veins and stockworks also contain significant Au. Both the bedding-parallel and leader veins are deformed, leading

Windh (1995) to suggest a pre- to syn-tectonic timing, relative to the formation of the Hill End Anticline, of veins and Au mineralisation.

In the Sofala-Wattle Flat area, rocks of the Late Ordovician (~450 Ma) Sofala Volcanics have been thrust over younger siliciclastic rocks of the Chesleigh Formation. The Au is hosted by both stockwork vein systems and by laminated quartz veins that contain arsenopyrite, pyrite, galena, chalcopyrite and gold. These veins are hosted by the Sofala Volcanics, a Devonian dyke that intrudes the Sofala Volcanics and the Chesleigh Formation in the structural footwall of the Spring Gully Fault, one of the thrust faults juxtaposing the Late Ordovician over Late Silurian Rocks (Rowley, 1994; Downes et al., 2008). In the Sofala Volcanics, the auriferous veins are associated with carbonate-rich alteration assemblages (Rowley, 1994).

Although the Gulgong area was dominated by Tertiary placers, lode gold deposits, from which the placer deposits were likely derived, are hosted by latite and volcanoclastic sandstone of the Late Ordovician (460–445 Ma: Meakin and Morgan, 1999) Burranah Formation and by monzodiorite bodies (~435 Ma: Black, 1998) that intrude this unit (J Watkins, unpub data, in Downes et al., 2008). The Au is hosted by stockwork quartz veins associated with propylitic and sericitic alteration assemblages. The main sulphide minerals in the veins include arsenopyrite, pyrite and chalcopyrite. Gold is present as free gold and as solid solution in pyrite and arsenopyrite (J Watkins, unpub data, in Downes et al., 2008).

Although limited, age constraints on geological events in and around the Hill End Trough suggest a complicated history, with Au mineralisation potentially spanning most of it. The earliest post-depositional event was a major east-west compression that resulted in development of the north-trending anticline and a strong cleavage. This cleavage pre-dated regional greenschist (biotite grade) metamorphism (Windh, 1995). Lu et al. (1996) presented a ^{40}Ar - ^{39}Ar plateau and four total fusion ages of hydrothermal biotite that gave consistent ages of ~360 Ma, which they interpreted as the age of regional metamorphism. Lu et al. (1996) interpreted a complex age spectrum from muscovite associated with a structurally early vein to indicate initial vein emplacement at 380–370 Ma, followed by recrystallisation associated with a younger fluid flow event at ~345 Ma. The earlier age corresponds with the Tabberabberan Orogeny, whereas the latter event corresponds to the Kanimblan Orogeny (see below).

Data on the timing of Au mineralisation, although not definitive, may suggest multiple mineralising events. The most robust age data, ^{40}Ar - ^{39}Ar plateau ages from muscovite associated with two phases of Au mineralisation at the Hill End cluster, yielded ages of ~358 Ma for the earliest phase and ~343 Ma for the later phase (Lu et al., 1996). These data span the Kanimblan Orogeny (see below), and the younger age is similar to the inferred younger fluid flow event described earlier. However, Pb isotope model ages, calculated from the Lachlan model of Carr et al. (1995), yielded a range of ages. In the Hill End cluster, Downes et al. (2008) reported model ages from 420 to 320 Ma, with most between 420 and 350 Ma. Moreover, three analyses of Pb-rich phases from the Hargraves deposit clustered around 420–410 Ma, and a number of other deposits (including those from the Sofala-Wattle Flat cluster) had model ages of 380–370 Ma (Downes et al., 2008). This suggests that like the Victoria goldfields, the Hill End Trough has experienced multiple Au events, in this case corresponding to the Bindian (420–400 Ma), Tabberabberan (380–370 Ma) and Kanimblan (360–345 Ma) orogenies. This scenario is consistent with the structural observations of Windh (1995), who demonstrated protracted vein development, with the earliest veins predating folding and the latest veins post-dating cleavage development.

The Wyoming deposit, which is hosted by a volcanoclastic sequence that forms part of the Ordovician Goonumbla Volcanics, 50 km north of the Northparkes porphyry Cu-Au deposits (Fig. 59), consists of quartz±carbonate±albite±pyrite±arsenopyrite veins and breccias. Most of the ores are hosted by feldspar-phryic andesite bodies that have intruded the volcanoclastic rocks and carbonaceous mudstones (Chalmers et al., 2007a). The deposit comprises several discrete lenses of which Wyoming One and Wyoming Three are the largest, making up a resource of 7.13 Mt grading 2.70 g/t Au (www.alkane.com.au [accessed 21 August 2008]). An association of the veins with sericite-carbonate-albite-quartz±chlorite±pyrite±arsenopyrite alteration assemblages led Chalmers et al. (2007a), in combination

with other data, to conclude that the Wyoming deposits were lode Au deposits. Although no definitive geochronology data were available, these authors inferred a late-Tabberabberan (400-380 Ma) age for mineralisation.

Chalmers et al. (2007a) suggest that the London-Victoria Au deposits 60 km to the south of Wyoming had a similar origin to the Wyoming deposit, and Chalmers et al. (2007b) suggested that most of the Au-only districts in central and south central New South Wales, including Adelong (~390 Ma: Perkins et al., 1995), Lucknow, West Wyalong, Parkes, Forbes, Young, Bodangora and Stuart Town as well as the Hill End and Gulgong districts described above, contained lode gold deposits or associated alluvial systems. In total these districts have produced over 114 t of Au (Chalmers et al., 2007b), and when combined with the deposits of the Hill End Trough and the Wyoming deposits, lode gold deposits in New South Wales have collective global resources of 228 t Au. Moreover, these deposits are widespread (Fig. 59). Although limited age data and geological relationships suggest a Tabberabberan age (400-365 Ma) for at least some of these deposits, more data are required to understand the geodynamic environment in which these deposits formed and the relationship to lode gold deposits further south in Victoria and Tasmania.

3.3.9. Epigenetic Cu-Au and Zn-Pb-Ag deposits, Cobar Trough and Girilambone district

The extensional Cobar Basin in west-central New South Wales (Fig. 59) is comprised predominantly of turbiditic siliciclastic rocks with two phases of basin fill. The older phase, the Nurri Group, consists of an upwards-fining sequence that fills the eastern part of the basin, whereas the younger phase, the Amphitheatre Group, fills the western part of the basin and contains two cycles, a lower upwards-coarsening sequence and an upper, finer-grained, thinly-bedded sequence (Glen et al., 1996). The age of these units is constrained between ~420 Ma, the age of Silurian granites the basin unconformably overlies (Pogson and Hillyard, 1981), and ~400 Ma, the age of a low grade cleavage-forming event (Glen et al., 1992).

Both units are mineralised, with the Amphitheatre Group containing both the Endeavour (aka Elura) and CSA deposits, and the Nurri Group containing the deposits near the town of Cobar (e.g., Great Cobar, New Occidental and Chesney). Global resources for the Cobar area total 4.20 Mt Zn, 2.47 Mt Pb, 1.76 Mt Cu, 4940 t Ag and 139 t Au (based mostly on Stegman [2001] with additional data on production from the Mount Boppy and minor deposits from Stegman and Stegman [1996]). Major deposits include the Endeavour (42 Mt grading 8.6% Zn, 5.4% Pb and 96 g/t Ag), CSA (48 Mt grading 1.1% Zn, 0.3% Pb, 3.1% Cu and 18 g/t Ag) and Peak (5.2 Mt grading 1.0% Zn, 1.1% Pb, 0.8% Cu, 8.4 g/t Ag and 9.1 g/t Au) deposits. These deposits illustrate the diverse metallogeny of the Cobar ores: the deposits range from Zn-Pb-Ag-rich through to Cu-rich through to Au-rich. Although overall individual ore deposits are characterised by specific metal assemblages, all well-described deposits (CSA, Peak and Endeavour: Cook et al., 1998; Shi and Reed, 1998; Webster and Lutherborrow, 1998) contain zones characterised by the other metal assemblages. Stegman (2001) indicated a broad stratigraphic zonation, with Au-rich deposits localised in the stratigraphically lowermost Chesney Formation (Nurri Group), the Cu-Au deposits hosted by the Great Cobar Slate (uppermost Nurri Group), and the Zn-rich deposits hosted by the CSA Formation (lowermost Amphitheatre Group). In addition, he indicated that the deposits are Au-rich immediately south of the town of Cobar and become more Cu-rich and Au-poor and then Zn-rich to the north and south.

These deposits also share a close association with structural elements. The Peak and CSA deposits are localised within a zone of high strain along the eastern margin of the Cobar Basin and, in detail, the lodes crosscut bedding and are commonly parallel to a well developed west-dipping cleavage (Cook et al., 1998; Shi and Reed, 1998; Lawrie and Hinman, 1998). The Endeavour deposits consist of a series of sub-vertical pipe-like bodies that are localised within the hinge zone of an anticline (Lawrie and Hinman, 1998; Webster and Lutherborrow, 1998). In the CSA and Endeavour deposits, the ores include massive to semi-massive sulphide, siliceous ore and stringer ore (Shi and Reed, 1998; Webster and Lutherborrow, 1998). In contrast, the Peak deposit is dominated by sheeted quartz veins with only minor semi-massive sulphide

(Cook et al., 1998). Mineralogically, the deposits contain similar ore minerals, although the proportion is variable. Typically the ores contain pyrite and pyrrhotite in variable ratios, along with variable quantities of chalcopyrite, sphalerite and galena and minor to trace tetrahedrite-tennantite, enargite, arsenopyrite, magnetite, Bi minerals and electrum (Lawrie and Hinman, 1998). Stegman (2001) reported an early magnetite-scheelite-wolframite stage. At Endeavour, the ores are zoned, with a pyrrhotite-rich cores grading outward and upward into pyrite-rich peripheral zones (Schmidt, 1990; Lawrie and Hinman, 1998). Alteration zones associated with the deposits include silicified and chlorite-rich zones. The latter assemblage is only weakly developed at Endeavour, though more extensively developed in the Cu- and Au-rich deposits (Cook et al., 1998; Shi and Reed, 1998; Lawrie and Hinman, 1998).

Structural relationships suggest that the vast majority of mineralisation occurred contemporaneous with the development of a regional penetrative cleavage interpreted by Glen et al. (1992), using ^{40}Ar - ^{39}Ar methods, to have been developed at 400-395 Ma. This interpretation was supported by ^{40}Ar - ^{39}Ar results from Au-related muscovite from the Peak deposit, which yielded an age of ~402 Ma (Perkins et al., 1994). However, muscovite from Zn-Pb-rich ore and altered volcanic rocks yielded ages of ~385 Ma and ~383 Ma, respectively. Based on these results, Perkins et al. (1994) interpreted Au- and Cu-rich mineralisation at the Peak and CSA mines to have formed at 405-400 Ma, contemporaneously with regional fabric formation associated with inversion of the basin (during the Bindian Orogeny), whereas Zn-rich mineralisation occurred during normal faulting associated with relaxation after this shortening event. These results are perhaps supported by Pb isotope data that indicated that the Zn-rich Endeavour (Elura) and CSA deposits are more radiogenic (and possibly younger) than the Cu-Au-rich New Cobar, Peak and Wood Duck deposits (Laurie and Hinman, 1998).

However, Sun et al. (2000) reinterpreted the previous ^{40}Ar - ^{39}Ar data to suggest that inversion of the Cobar Basin occurred at 389-385 Ma, with synchronous mineralisation and a possible young stage of mineralisation at 379-376 Ma. In this case, deformation and mineralisation would be related to the early phases of the Tabberabberan Orogeny. Under either interpretation, mineralisation is inferred to have occurred during the Tabberabberan Cycle.

Approximately 100 km to the south-southeast of Cobar, the Hera deposit has an initial resource of 2.2 Mt grading 3.4 g/t Au, 4.2% Zn, 3.1% Pb, 0.2% Cu and 18 g/t Ag (Jones and MacKenzie, 2007). This deposit, which was discovered in 2000 at a depth of 300 m below the surface (Collins et al., 2004), consists of narrow lenses of pyrrhotite-sphalerite-galena-pyrite veins that are hosted by silicified and chloritised sandstone and siltstone (Skirda and David, 2003) that is part of the Cobar Basin.

Copper deposits of the Girilambone district, approximately 100 km to the east-northeast of Cobar, are hosted by the Girilambone Group of Ordovician age that underlies the Cobar Basin. This district had pre-mining resources totalling 0.58 Mt Cu (data from Fogarty, 1998; Erceg and Hooper, 2007), with the majority of this resource residing in the Tritton deposit (14 Mt grading 2.7% Cu, 12 g/t Ag and 0.3 g/t Au: Fogarty, 1998). Until recently, mining had concentrated on SX-EW leachable near-surface deposits, although mining commenced on the primary Tritton deposit in 2004.

The deposits are hosted by the Tritton Formation, which consists of arenite, greywacke and slate. Most of the previously mined deposits consist of Cu carbonate and supergene enriched sulphide zones. These supergene zones are developed upon massive and stringer pyrite-chalcopyrite lenses, the richest and largest of which is the Tritton deposit (Fogarty, 1998). This deposit consists of two lenticular ore lenses hosted by a sequence of quartz sandstone and mica schist. The stratigraphically lower ore lens consists of massive to banded pyrite-chalcopyrite that overlies carbonate-epidote-magnetite altered mafic schist (Fogarty, 1998; Erceg and Hooper, 2007). The ores appear to be vertically zoned, as follows (from lower to upper zones): (1) pyrrhotite-pyrite → pyrite → pyrite-arsenopyrite, and (2) chalcopyrite → chalcopyrite-bornite → tennantite-chalcopyrite (T Leach in Erceg and Hooper, 2007). Alteration assemblages grade outwards from the ores as follows: quartz-magnetite-carbonate → stilpnomelane-quartz-biotite±magnetite → chlorite-carbonate±biotite±quartz±sulphides (Erceg and Hooper, 2007). In contrast to earlier interpretations that the Girilambone deposits were Besshi-type VHMS deposits, Erceg and Hooper (2007) interpreted them as

epigenetic with similar characteristics to the deposits of the Cobar district. If so, the Girilambone deposits most likely formed during or shortly after contraction associated with the Tabberabberan (or Bindian) Orogeny.

3.3.10. Lode gold deposits, Tasmania (with contributions from R Bottrill)

Based on data compiled in Bottrill et al. (1992) and Seymour et al. (2006), Ordovician to Early Devonian siliciclastic rocks in the Beaconsfield area and of the Mathinna Supergroup (Fig. 60) have total resources and production of just over 100 t Au, with major groups of deposits at Beaconsfield (3.25 Mt grading 19.0 g/t Au for 62 t), in the Lyndhurst-Mangana corridor (18 t) and in the Denison-Lisle (10 t) and Lefroy-Back Creek goldfields (10 t). With the exception of the Beaconsfield deposit, the eastern Tasmanian lode gold deposits are small. However, all deposits are generally high grade (10-20 g/t), with many deposits averaging over one oz/t (31 g/t) (Bottrill et al., 1992).

The smaller Tasmania deposits are located in the East Tasmania Terrane, hosted by turbiditic sedimentary rocks of the Ordovician to Lower Devonian Mathinna Supergroup and, to a lesser extent, by granites. The sediment-hosted deposits are generally within a few kilometres of Tabberabberan-aged granites. In the East Tasmania Terrane these granites range in age from 400 to 378 Ma, becoming increasing younger to the west (youngest ages in western Tasmania are ~351 Ma: Black et al., 2005). In the Lyndhurst-Mangana gold corridor a series of goldfields are hosted by a north-northwest-trending, 80 km by 10-20 km belt of turbiditic sedimentary rocks of the Mathinna Supergroup between the 400-378 Ma (Black et al., 2005) Blue Tier and 390-386 Ma (Black et al., 2005) Scottsdale Batholiths. The deposits are more narrowly focussed, forming a 2-km-wide belt, with high-grade ore shoots limited to a few 10s of metres in width (Edwards, 1953). Bottrill et al. (1992) indicated that the deposits are hosted by quartz veins and breccias with minor sulphides dominated by pyrite and arsenopyrite with lesser chalcopyrite, galena and sphalerite. The veins range in thickness between a few centimetres to eight metres, with strike lengths of up to 2 km (Finucane, 1935). Structurally the veins are commonly bedding-parallel and have a steep dip (Bottrill et al., 1992); in some locations these veins have been folded. ^{40}Ar - ^{39}Ar data from vein-related sericite from the Mathinna and Mangana goldfields yielded ages of 394-390 Ma (Bierlein et al., 2005), which overlap the ages of the nearby granite batholiths.

The Denison-Lisle goldfield (Fig. 60C) has two main parts, the Denison goldfield in the north and the Lisle-Golconda goldfield in the south. The Denison goldfield is very similar to the Lefroy gold field (see below), with steeply dipping quartz lodes with patchy gold and sulphide rich shoots hosted in Mathinna Beds, and probably 2-3 km above a granodiorite (Leaman and Richardson, 1992).

About 10t of alluvial gold was produced in the Lisle-Golconda goldfields to the south (Bottrill, 1994) and this was apparently derived from numerous small but rich sulphide veins, stockworks and disseminations present in both hornfelsed Mathinna Beds and granodiorite (Callaghan, in Taheri et al., 2004). These deposits are interpreted as intrusion-related type deposits, coeval with the granodiorite bodies which are probably small plutons related to the nearby Scottsdale Batholith. The veins are mostly quartz poor and rich in pyrite and arsenopyrite, with minor galena, sphalerite, bismuthinite, molybdenite and rare maldonite. These deposits are interpreted to have been formed at ~385 Ma based on ^{40}Ar - ^{39}Ar analyses of hydrothermal biotite from a monzodiorite dyke (Bierlein et al., 2005).

In the small Whiting prospect, in the nearby St. Patricks River area, gold occurs in arsenopyrite-rich pods in granodiorite/hornfels contact zones, with silver sulphosalts (pyrargyrite, polybasite and fribergite; Bottrill, 2005).

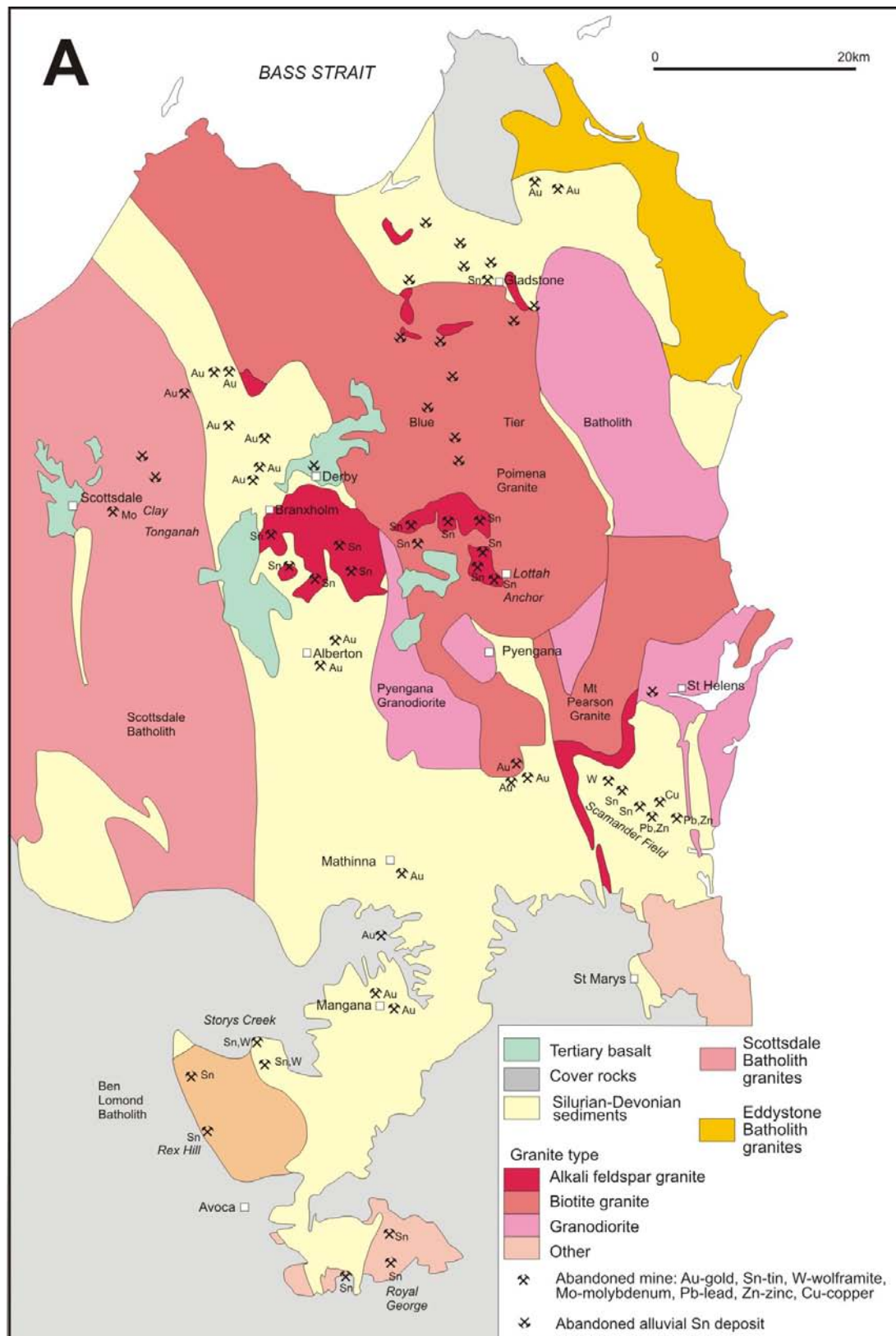


Figure 60. Distribution of lode gold, granite-related Sn-W and other deposits in Tasmania. A - Deposits in northeastern Tasmania. Figure modified from Solomon and Groves (2000).

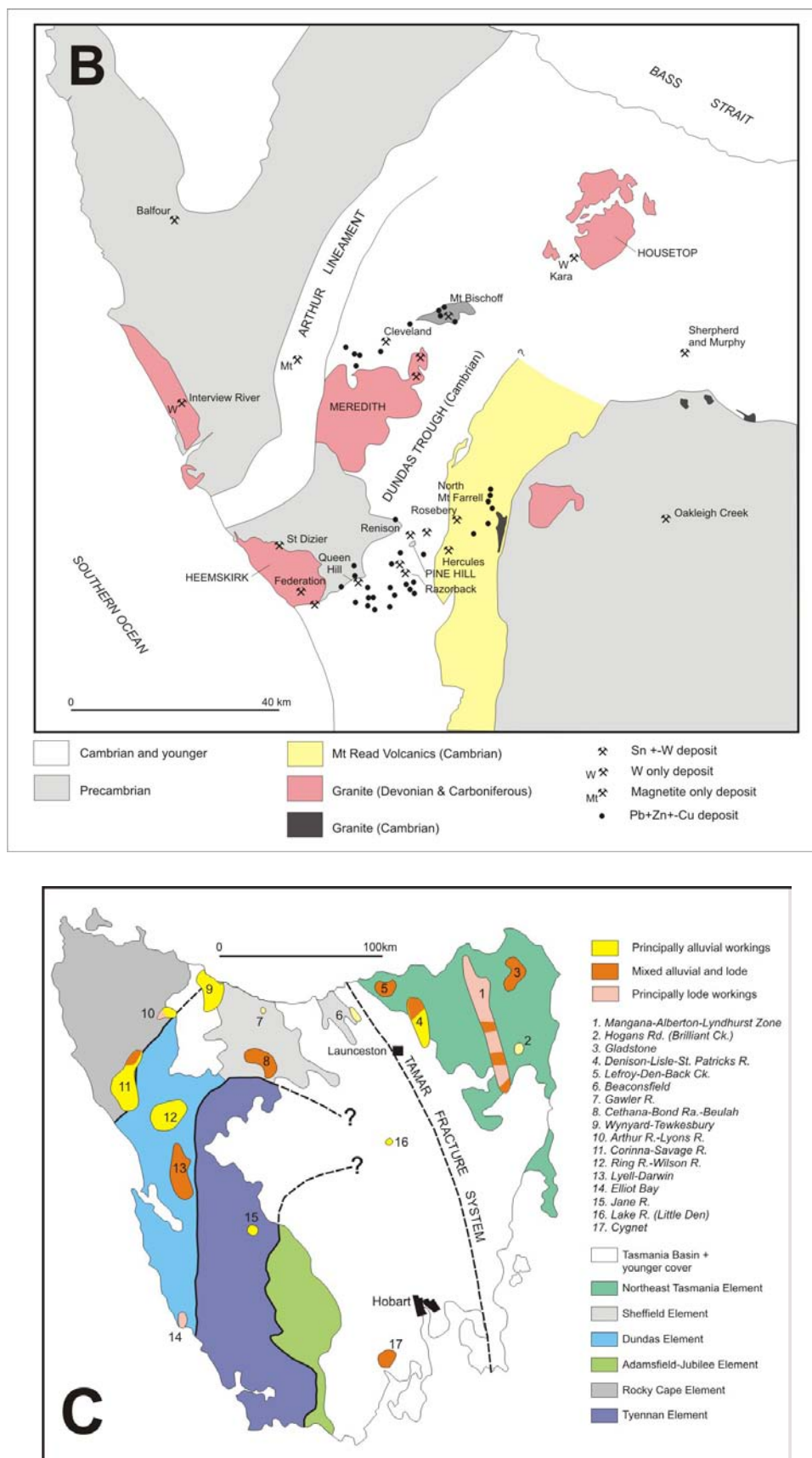


Figure 60. Distribution of lode gold, granite-related Sn-W and other deposits in Tasmania (continued). B – Sn-W and other deposits in western Tasmania. Figure modified from Solomon and Groves (2000). C. Gold fields of Tasmania. Goldfield locations from Bottrill et al. (1992), superimposed on the element map of Seymour and Calver (1995).

In the Lefroy and Back Creek goldfields, Au at Lefroy is hosted by broadly east-trending, sub-vertical quartz veins, parallel to thrust movements in Mathinna group siltstones and slates. They have been worked for up to 1220 m along strike, with gold and sulphides (arsenopyrite, chalcopyrite, boulangerite and stibnite) confined to structurally-controlled tabular shoots (to about 200m long) plunging at $\sim 45^\circ$ west (Reed, 2002; Van Moort and Russell, 2005). As with most Tasmanian reef deposits, near surface Au is present as free gold in weathered sulphidic veins, but with increasing depth auriferous pyrite becomes more prevalent. Based on gravity modelling, the deposits are within 2-3 km of buried granites (Leaman and Richardson, 1992).

In contrast to deposits of the East Tasmanian Terrane, the Tasmania reef at Beaconsfield is hosted by the Lower Ordovician Salisbury Hill and Eaglehawk Gully Formations, which locally comprise the Denison Group (Reed, 2001). The Salisbury Hill Formation (formerly lower Transition beds) consists of conglomerate and sandstone, whereas the Eagle Hawk Gully Formation (formerly upper Transition beds) consists mostly of calcareous siltstone, with minor limestone at the stratigraphic top (Hills, 1998). This sequence forms part of the Beaconsfield Block, a fault-bounded block between the Neoproterozoic Badger Head Block to the west and the East Tasmania Terrane to the east (Elliot et al., 1993). The Beaconsfield Block comprises Cambrian siliciclastics rocks (equivalent to the Port Sorell Formation) and the ultramafic rocks of the Anderson Creek Complex, Ordovician siliciclastic rocks and limestone (including the Salisbury Hill and Eaglehawk Gully Formations), and Silurian to Devonian mudstone and siltstone. Elliot et al. (1993) indicated that these rocks are tectonically imbricated along east-dipping thrusts, with this thrusting related to the Tabberabberan juxtaposition of the East and West Tasmania Terranes. If this interpretation is correct, the Beaconsfield deposit differs from smaller deposits further to the east in being in the footwall to the suture between these two terranes.

The Tasmania reef is hosted by a D_2 shear ($60^\circ \rightarrow 131^\circ$) that cuts the sub-parallel orientations of bedding and thrusting (D_1 : $60\text{--}65^\circ \rightarrow 047^\circ$) at a high angle, resulting in an ore panel that plunges steeply to the east (Hicks and Sheppy, 1990; Hills, 1998). The vein is largely restricted to the Salisbury Hill and Eaglehawk Gully Formation. Where the underlying Cabbage Tree Conglomerate or the overlying Flowery Gully Limestone are encountered, the vein becomes branched and rapidly dies, restricting the vein to a horizontal length of about 400 m (Bottrill et al., 1992). The geometry of the vein relative to the thrust planes suggests that it may have been formed in a local transtensional environment associated with oblique accretion (from the north-northeast) of the East Tasmania Terrane onto the West Tasmania Terrane.

The Tasmania reef averages about 3 m in width, with a maximum width of 5.4 m (Hicks and Sheppy, 1990; Hills, 1998). Bottrill et al. (1992) indicate that the vein is zoned with an ankerite-rich core, a gold-sulphide-enriched intermediate zone and quartz-rich margins. The sulphides include pyrite, arsenopyrite and chalcopyrite, with minor sphalerite, galena and tetrahedrite (Bottrill et al., 1992; Russell and van Moort, 1992).

In addition to its location in the West Tasmania Terrane, west of the Tamar Fracture, the Beaconsfield deposit also differs from the other deposits in lacking a close spatial association with granites (4-6 km from granite: Leaman and Richardson, 1992), in being spatially associated with a mafic-ultramafic complex, and in having a possibly older age. Although not well constrained, laser ablation $^{40}\text{Ar}\text{--}^{39}\text{Ar}$ analysis of ore-related fuchsite yielded a plateau age of 400 ± 5 Ma, compatible with a total fusion age from another Beaconsfield sample of ~ 405 Ma, apparently older than the $\sim 394\text{--}390$ Ma age of the Lyndhurst-Mangana gold corridor (Bierlein et al., 2005).

Although the most significant lode Au deposits in Tasmania are in the northeastern part of the State, Bottrill et al. (1992) also identified several goldfields of similar age in the west. The main fields in western

Tasmania are around Queenstown, Moina, Savage River-Corinna and Lakeside (Tullah). These contain mostly small but often very rich quartz and/or carbonate veins and stockworks, sometimes with very coarse gold. Lakeside contains arsenopyrite-rich quartz veins with anomalous tin and contains a resource of about 0.75Mt @ 2g/t. Numerous vein style deposits of Ag-Pb, Cu, Sb and As occur throughout Tasmania and many also carry significant gold and are probably related. ^{40}Ar - ^{39}Ar analyses by Bierlein et al. (2005) yielded an age of ~383 Ma based on a silicified pelitic rock from within a quartz-carbonate sericite-sulphide lode King River gold mine near Queenstown. This result suggests that the 400-380 Ma lode gold event may have been more widespread.

3.3.11. Mount Morgan and related deposits, Calliope Arc

Like many world-class ore deposits, the origin of the Devonian Mount Morgan deposit (Fig. 61), which produced a total of 50 Mt grading 0.72 % Cu and 4.99 g/t Au (Ulrich et al., 2002), is controversial. Over the last few decades, workers have advocated VHMS origins (e.g., Frets and Balde, 1975; Taube, 1986) as well as epigenetic origins related to the ~381 Ma (Golding et al., 1993) Mount Morgan Tonalite (Arnold and Sillitoe, 1989) and Permian intrusions (Cornelius, 1969). As first discussed by Lawrence (1967, 1972), the Mount Morgan ores have been thermally annealed by the intrusion of a magma, most likely the Mount Morgan Tonalite. The deposit is hosted by a roof pendant of thermally metamorphosed volcanoclastic, siliciclastic and carbonate rocks within this tonalite, which, at most, is several million years younger than the host sequence (Mine Sequence). The most recent study (Ulrich et al., 2002) advocated a hybrid origin involving deposition of barren massive sulphide at or near the seafloor, which is overprinted by quartz-chalcopyrite-pyrite stockwork sourced from the Mount Morgan Tonalite.

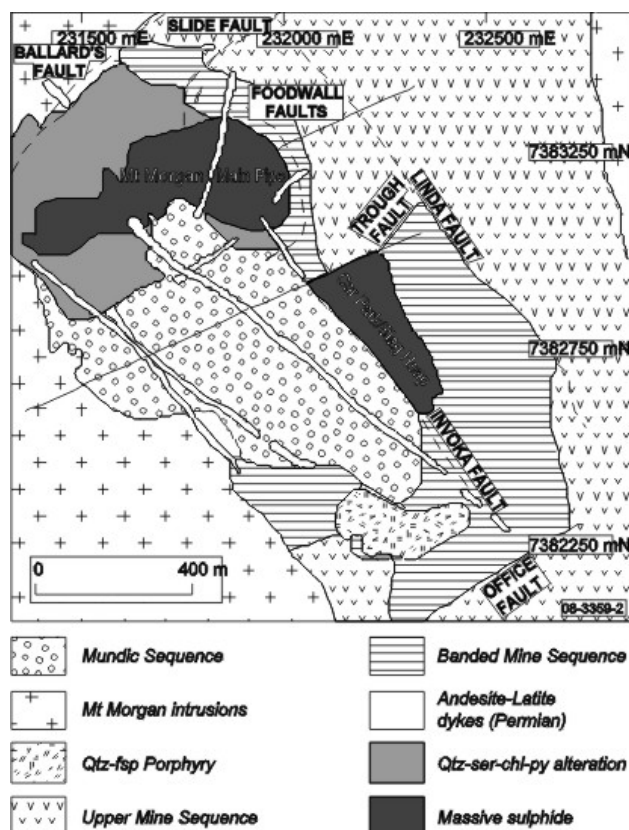


Figure 61. Geology of the Mount Morgan deposit (after Ulrich et al., 2002).

The Mount Morgan deposit consists of three sulphide bodies (Fig. 61), two of which, the Main Pipe and Sugarloaf orebodies, were mined up until 1982 and the third, the Car Park/Slag Heap zone, was discovered

brecciated, massive sugary pyrite with minor sphalerite and pyrrhotite that is overprinted by quartz-chalcopryrite stockwork. The Sugarloaf orebody comprises stringer and disseminated pyrrhotite, pyrite and chalcopryrite in highly siliceous quartz-sericite-chlorite rock, and is interpreted to be the stringer zone below the Main Pipe orebody, which either formed at or shallowly below the seafloor. The stratiform Car Park/Slag Heap zone is zoned, with an upper, thin magnetite-pyrite-chlorite assemblage overlying massive pyrite, which in turn overlies stringer pyrite. The Car Park/Slag Heap zone, which is somewhat elevated in Zn, but has much lower grade Cu and Au, does not have the quartz-pyrite-chalcopryrite stockwork that characterises the other two orebodies (Ulrich et al., 2002). The Car Park/Slag Heap zone is localised stratigraphically below the Main Pipe orebody (Fig. 61).

All three mineralised zones are associated with quartz-sericite-chlorite±pyrite altered zones, which can extend 100s of metres laterally and vertically below the zones. Around the Main Pipe orebody, the alteration assemblage is zoned, with an inner more siliceous zone (Golding et al., 1993; Messenger et al., 1998). However, Arnold and Sillitoe (1989) described local retrogressed amphibole±biotite assemblages and cited unpublished reports of garnet- and epidote-bearing assemblages. They interpreted these assemblages, which they described as widespread, though patchy, as early-formed contact alteration assemblages associated with the emplacement of the Mount Morgan Tonalite. Arnold and Sillitoe (1989) also note that marginal zones of the Mount Morgan Tonalite adjacent to the orebodies were silicified and pyritised, with local hydrothermal biotite and amphibole.

Ulrich et al. (2002) presented a model in which barren massive pyrite was formed at or near the seafloor from the circulation of (evolved) seawater. As the system evolved, magmatic-hydrothermal brines and vapors evolved from early phases of the Mount Morgan Tonalite were incorporated into the hydrothermal system, producing the overprinting quartz-chalcopryrite stockwork veins.

The equivalents to the Mount Morgan Mine Sequence, the Capella Creek beds also contain small apparently stratiform base metals deposits, the best described of which is the Ajax deposit. This deposit is hosted by rhyolitic tuffaceous rocks that have been altered to a pyritic quartz-sericite assemblage. The mineralised rock typically assays 0.5% Cu and 2.5% Zn with minor Pb, Ag and Au (Large, 1980). This deposit is most likely a VHMS deposit.

3.3.12. Mineral potential

Synthesis and analysis of geological and metallogenic data suggest that the Tabberabberan cycle was characterised by five major geodynamic systems, in order of decreasing age:

1. North Queensland arc-backarc system;
2. East Lachlan arc-backarc system;
3. Bindian Orogeny;
4. Gamilaroi-Calliope arc-backarc system; and
5. Tabberabberan Orogeny.

Unlike the Benambran cycle, knowledge of these tectonic systems is reasonably good, although in detail different models exist to account for some features. Based on existing data, predictions are made below about the existence and likely extent of mineral systems for each of these tectonic systems (Figs. 63 and 64).

3.3.12.1. North Queensland arc-backarc system

In north Queensland, the dominant tectonic system involved deposition of turbidite with lesser tholeiitic basalt and limestone between the Early Silurian and Early Carboniferous in the Hodgkinson Province. This deposition is linked in time with the emplacement of both I- and S-type granites, some of which have arc-like affinities, in the Pama Province that extends around the Hodgkinson and Broken River provinces from Charters Towers in the south to Cape York in the north. These rocks were affected by a thermotectonic

event, which temporally corresponds with the Bindian Orogeny, at 410-400 Ma (see [section 3.3.12.3](#)), which coincided with a hiatus in sedimentation in the Graveyard Creek Subprovince and a change in sedimentation style in the Hodgkinson Province. Sedimentation in north Queensland appears to have continued through the Tabberabberan Orogeny ([section 1.2.4](#)).

Two broad tectonic models have been proposed for this tectonic system, both of which infer west-dipping subduction. In the first model (e.g., Henderson, 1987), the Hodgkinson and Broken River Provinces are interpreted as fore-arc basins with the Pama Province as the arc. Alternatively, the Hodgkinson Province is interpreted as a backarc (Arnold and Fawcner, 1980). Determining confidently the likely tectonic model has important implications for mineralisation. If the Hodgkinson Province is forearc, the Pama Province is likely the associated magmatic arc and the backarc would be located further inland. In such a situation, the granites of the Pama Province might be associated with porphyry and epithermal mineral systems, although subsequent erosion may have removed such deposits. In addition, there would be potential for backarc-related mineral (e.g., VHMS) systems further inboard to the west. If the Hodgkinson Province is a backarc, however, the Pama Province may have potential for intrusion-related Sn, W and Mo deposits. Unfortunately, the Pama Province is not associated with significant mineralisation, although the presence of VHMS deposits in the Hodgkinson Province is more consistent with a backarc setting for this province.

3.3.12.2. East Lachlan arc-backarc system

In the Lachlan Orogen, deposits with ages between the Benambran and Bindian Orogenies (i.e. 435-410 Ma) appear to have a well defined spatial pattern, with intrusion-related gold deposits near the coast (e.g., Major Creek), VHMS deposits hosted by extensional basins in the East Lachlan (e.g., deposits hosted by the Goulburn, Cowombat and, possibly, Buchan rifts) and intrusion-related Sn, W, Mo and Cu-Au deposits associated mainly with S- and, to a lesser extent, I-type granites of the Wagga Sn Belt of the Central Lachlan. As discussed in [section 2.4](#), our favoured interpretation of this system is that it was associated with extension related to slab roll-back (e.g., Collins and Richards, 2008), with the subduction zone offshore (present day) to the east (c.f. [Fig. 37](#)), with intrusions such as parts of the Bega Batholith perhaps a continuation of the Calliope-Gamilaroi Arc. [Figure 62](#) shows schematically this tectonic system in section based on the tectonic model of Collins and Richards (2008) for this period. Magmatic-related deposits near the coast are interpreted to be located just inboard of the remnant volcanic arc, VHMS deposits are localised in back-arc basins, and the Sn, W and Mo deposits are associated with crustal melts that formed inboard of the back-arc basin as a consequence of decompression melting of the lower crust. Based on this interpretation, the Central and Eastern Lachlan would have potential for both VHMS and intrusion-related deposits. Although not shown in [Figure 63](#), this potential may be zoned, with porphyry Cu-Au, intrusion-related Au and, possibly, epithermal potential highest in the east, near the coast, potential for VHMS and related deposits highest in extensional basins such as the Goulburn Basin, and potential for Sn, W and Mo deposits highest in the Wagga Sn belt, furthest inland. Emplacement of these granites into older mafic-ultramafic belts such as the Fifield Complex raises the possibility of Auebury-type hydrothermal Ni deposits.

Although the metallogenic implications of our favoured tectonic model are presented above, other models exist, particularly those invoking multiple subduction zones ([section 2.4](#)). For example, Gray and Foster (1997, 2004) presented a model ([Fig. 23](#); see [section 2.4](#)) invoking bivergent subduction, with west directed subduction underneath the West Lachlan and east-directed subduction below the Central and Eastern Lachlan, with consumption of the intervening oceanic crust resulting in juxtaposition of the Central and West Lachlan. This model would predict the presence of arc-related (e.g., porphyry and epithermal) mineralisation on both sides of the resultant structure (i.e. southwest margin of Central Lachlan and northeast margin of Western Lachlan) as well as backarc-related mineralisation (e.g., VHMS) further inboard from the margins. The presence of VHMS deposits in the Goulburn and related extensional basins is consistent with both this model and our favoured model of east dipping subduction along the eastern margin of the East Lachlan.

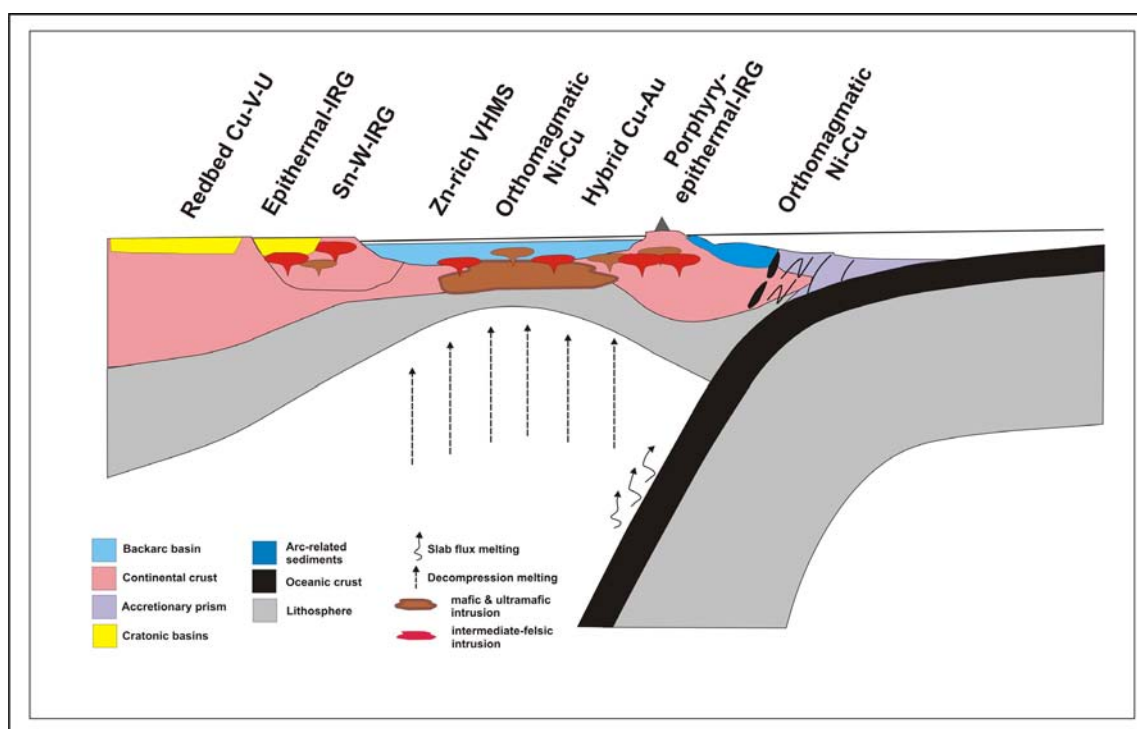


Figure 62. Schematic diagram, based on tectonic model of Collins and Richards (2008), showing the location of different types of mineral deposits.

3.3.12.3. Bindian Orogeny

Although Bindian Orogeny-related gold appears to be a minor part of the Victorian gold fields and an uncertain part of the lode Au deposits in New South Wales, it is the major event in northern Queensland and Tasmania, producing the Charters Towers goldfield (~407 Ma) and smaller goldfields in the Etheridge Province, and the Tasmania reef at Beaconsfield (405-400 Ma) in northern Queensland, gold deposition overlaps in time with Pama Province granites, although Kruezer (2005) inferred that deposits in the Charters Towers Goldfield were orogenic, and not intrusion-related, in origin. The Tasmania lode is significantly removed from granites based on gravity data. Based on this distribution we consider that rocks affected by the Bindian Orogeny all have potential for lode Au deposits (Fig. 64), although these rocks in the Thomson Orogen have the highest potential, particularly in the vicinity of Charters Towers. Inversion of basins formed during the Bindian cycle during the Bindian Orogeny also may produce Cobar-type Cu-Au and/or Zn-Pb-Ag deposits (Fig. 64).

3.3.12.4. Gamilaroi-Calliope arc

As discussed in section 1.3.3, the Late Silurian to Early Devonian succession in both the southern (Gamilaroi Terrane) and northern (Calliope Arc) New England Orogen, are dominated by volcanic rocks and associated volcanoclastic rocks with associated sub-volcanic intrusions. Geochemically, the magmatic rocks of both terranes suggest island arc affinities (Offler and Gamble, 2002; Murray and Blake, 2005; Offler and Gamble, 2002). Exposure of these rocks is very limited, largely localised within anticlinorial zones that expose these rocks through younger cover of the Yarrol (northern New England Orogen) and Tamworth (southern New England Orogen) Terranes. In addition, the Gamilaroi-Calliope arc probably extends below the Clarence-Moreton Basin between the two parts of the New England Orogen.

Based on this tectonic setting, the presence of the Mount Morgan Cu-Au deposit, and the presence of extensive cover, the Gamilaroi-Calliope arc must be considered to have high potential for deposits of the porphyry-epithermal mineral system as well as for hybrid, Cu-Au-rich VHMS deposits (Fig. 63). In

addition, although not known, it is likely that a backarc basin would have been present inboard of this arc, with potential for Zn-rich VHMS deposits. The main impediment to exploration for these targets is the lack of exposure of the Gamilaroi-Calliope arc and the thickness of overlying rocks.

3.3.12.5. Tabberabberan Orogeny

The Tabberabberan Orogeny, which mostly involved broadly east-west contraction, affected the vast majority of eastern Australia, including the vast majority of Lachlan, Thomson and New England Orogen rocks older than 380 Ma. This orogeny is interpreted as the amalgamation and cratonisation of the Lachlan Orogen and occurred between 390 and 380 Ma ([section 2.4](#)). Despite the wide extent of this orogeny, known mineral production associated with it is relatively restricted, limited to an unknown proportion of lode gold in New South Wales, small, though rich lode gold deposits in the East Tasmania Terrane, and epigenetic Cu-Au and Zn-Pb-Ag deposits in the Cobar and Girilambone areas of New South Wales. Tabberabberan-aged lode gold has not been documented in the Victorian goldfields.

Despite the relatively limited distribution of Tabberabberan-aged lode gold deposits, this orogeny must be considered to have low to moderate potential for this type of deposits, with the best potential in central to southeastern New South Wales ([Fig. 64](#)). There is also untested potential for these deposits in covered regions of south-central Queensland. We do not consider the New England Orogen to have significant potential for these deposits as the Tabberabberan Orogeny is only weakly developed in those rocks.

The other deposit types that appear to be associated with the Tabberabberan Orogeny are epigenetic Cu-Au and Zn-Pb-Ag-(Cu-Au) deposits that are associated with, or shortly post-dated, the inversion of the Cobar basin between 390 and 375 Ma. By analogy with the Cobar and related Girilambone mineral systems, we consider that extensional basins of the Bindian-Tabberabberan Cycle in the Eastern and Central Lachlan in New South Wales and Victoria and basins of uncertain origin, such as the Hodgkinson in north Queensland have potential for epigenetic base metal deposits in the vicinity of structures related to Tabberabberan inversion. Potential for similar types of mineralisation may be present in turbiditic sequences in Tasmania ([Fig. 64](#)).

3.3.12.5. Other mineral systems

In addition to the potential described above, we also consider that the Bindian-Tabberabberan Cycle has potential for Irish-style or Mississippi Valley-type Zn-Pb deposits associated with carbonate rocks in the Hodgkinson and Broken River Provinces of north Queensland, and for intrusion-related Sn-W and Mo deposits associated with S-type granites in the Western Lachlan of western Victoria.

3.4. Kanimblan cycle (380-350 Ma)

The Kanimblan Cycle was marked by I-, S- and A-type magmatism associated with extension and rifting possibly related to a volcanic arc and subduction zone well off to the east. Deposits formed during this cycle temporally and spatially overlap this magmatism and include vein-hosted Au deposits in Victoria and granite-related Sn and W deposits in Tasmania. In addition, as discussed in [section 3.3.7](#), a significant proportion of lode gold deposits in New South Wales may be associated with the Kanimblan Orogeny.

3.4.1. Lode gold deposits, Victorian goldfields (380-365 Ma deposits)

After the ~440 Ma event, the 380-365 Ma hydrothermal event produced the largest quantity of Au in the Victorian goldfields. Deposits of this age are most prevalent the central and eastern part of the Victorian goldfields ([Fig. 54](#)), including the Costerfield Au-Sb domain of Phillips et al. (2003), the Woods Point and Walhalla goldfields in the very eastern part of the Melbourne Zone, and the Fosterville deposit in the northeastern Bendigo Zone. Gold of this age is also present at the small (0.22 t) Linton goldfield, which is

located in the Stawell Zone, southeast of Ballarat (Bierlein et al., 2001a), and unusual Au-Te-Cu-Sb-Bi-rich ores from the Day Dawn vein (Maldon) are also of this age (Bierlein et al., 2001a).

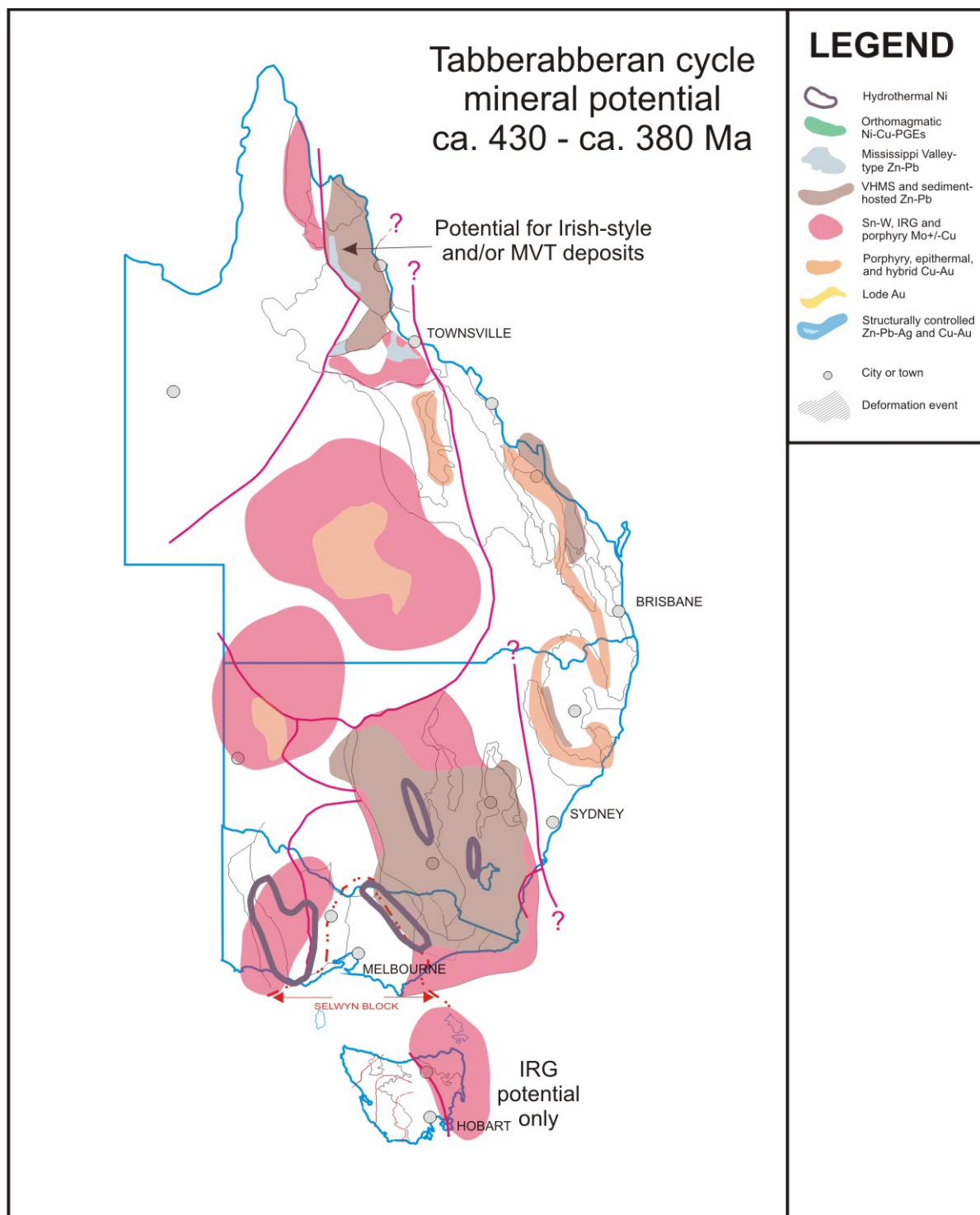


Figure 63. Mineral potential of the Bindian-Tabberabberan cycle.

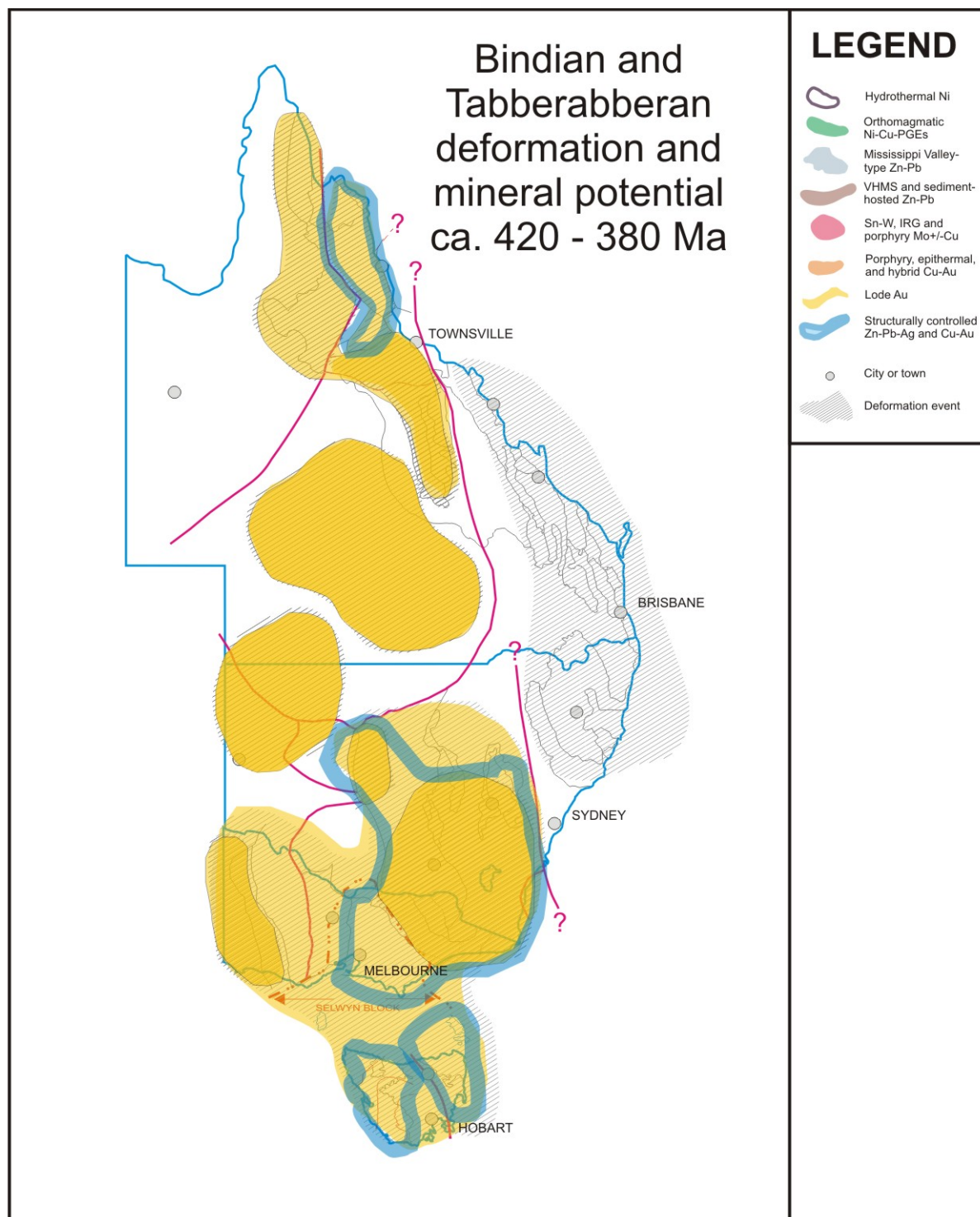


Figure 64. Mineral potential of the Bindian and Tabberabberan Orogeny.

The most significant deposits of this age are from the Melbourne Zone, which is dominated by the Walhalla (68 t) and Woods Point (52 t; Phillips and Hughes, 1998) goldfields in the far eastern part of the zone. In both of these goldfields, mineralisation is associated with, and overprints, the Woods Point Dyke Swarm, which comprises dykes ranging in composition from gabbro to rhyolite. These dykes intrude Late Ordovician to Early Devonian metasedimentary rocks (Bierlein et al., 2001a). The ores were hosted by quartz veins that form conjugate sets (ladder veins) in the Woods Point goldfield and laminated quartz veins in early faults that cut dykes in the Walhalla goldfield. The gold is associated with pyrite, arsenopyrite, galena, chalcopyrite and sphalerite in both goldfields and with minor bournonite, tetrahedrite

and jamesonite in the Woods Point goldfield (Phillips et al., 2003). The age of these deposits is constrained both by the age of the Woods Point Dyke Swarm and the age of sericite associated with the auriferous veins. Foster and Fanning (unpublished zircon U-Pb and hornblende ^{40}Ar - ^{39}Ar data, cited in Phillips et al., 2003) indicated that these rocks were emplaced at 378-376 Ma, which corresponds to the timing of Au-related sericite established using ^{40}Ar - ^{39}Ar analyses from the Walhalla and Great Rand deposits (Foster et al., 1998).

The Costerfield Au-Sb domain, which is characterised by the presence of stibnite as an important ore mineral in addition to gold, consists mostly of small goldfields (<10 t production) largely in the Melbourne Zone, but extending westward into the Bendigo Zone (Phillips et al., 2003). This domain overlaps the Central Victorian Magmatic Province (Campbell et al., 2003), a province characterised by relatively young (370-360 Ma), Ba-rich and reduced granites (both I- and S-type) with associated volcanic rocks (Rossiter, 2003). The deposits comprise quartz veins with high-level characteristics that cut siliciclastic rocks of the Castlemaine and Murrindindi Groups, and, to a lesser extent, volcanic rocks, red beds and granites of Late Devonian age (Campbell et al. (2003). The veins contain major quantities of stibnite, pyrite and arsenopyrite as well as minor to trace aurostibite, native antimony, berthierite, chalcostibite, chalcopyrite, galena, sphalerite, bournonite, boulangerite, pyrrhotite and marcasite (Campbell et al., 2003). In the Costerfield goldfield, the veins are aligned within the axis of the north-trending Costerfield Anticline. This goldfield produced an estimated 100,000 tonnes Sb but only 2.3 t Au (though at a grade of 15 g/t Au) (Campbell et al., 2003). However, most production in the Costerfield Au-Sb domain came from deposits for which stibnite is only present in trace amounts.

The Fosterville goldfield, which had global resources in excess of 45 t Au, differs from other Au deposits in the Victorian goldfields in that the Au is not hosted by thick quartz veins, but is disseminated through metasedimentary rocks and quartz-phyric rhyolitic dykes. The ores are associated with the north-trending Fosterville Fault, particularly adjacent to the pyrobitumen-bearing black shales, and contain disseminated pyrite, with lesser arsenopyrite and trace stibnite (Phillips et al., 2003). SHRIMP analyses of two samples of the rhyolitic dykes yielded imprecise ages of 388 ± 14 Ma and 394 ± 17 Ma (Bierlein et al., 1999, 2001b), providing a maximum constraint on the age of mineralisation. The best constraint on the age of mineralisation is ~ 381 Ma from ^{40}Ar - ^{39}Ar analysis of an altered rhyolitic dyke collected at surface (Bierlein et al., 2001a).

Although most Au produced from the Maldon goldfield (65 t) predated the emplacement of the ~ 370 Ma Harcourt granite (Phillips et al., 2003), the unusual Au-Te-Cu-Sb-Bi-rich ores from the Day Dawn vein are hosted by a diorite dyke with an age of 370 ± 4 Ma, which, although in a strict sense is a maximum age, is most likely the age of mineralisation (Bierlein et al., 2001a). In the nearby Eagleshawk-Linscott Reef, Ebsworth et al. (1998) recognised two stages of Au mineralisation: an early phase, in which the gold is associated with arsenopyrite and minor pyrite, galena, sphalerite, chalcopyrite and marcasite, and a later phase. The later phase occurs as seams within veins hosting the earlier phase or in hydraulic breccias overprinting the earlier assemblage, and comprises a gold-maldonite-joseite-native bismuth assemblage. This latter assemblage is probably equivalent to the assemblage at the Day Dawn vein.

Observations from the Walhalla, Woods Point, Costerfield, Fosterville and Maldon goldfields suggest that the 380-365 Ma mineralising event was quite diverse, ranging from high level epizonal Sb-Au deposits in the Costerfield Au-Sb goldfield, through the disseminated Fosterville deposit and the unusual Au-Te-Cu-Sb-Bi late stage assemblage of the Maldon goldfield, to the more typical Au-As assemblage of the veins in the Walhalla-Woods Point area. Importantly, all of these deposits are spatially and temporally associated with magmatism, suggesting that these deposits may have a significant magmatic-hydrothermal component. This differs from the the main, Late Ordovician (~ 440 Ma) Victorian Au event, in which Au mineralisation predates magmatism by 25-75 million years (Bierlein et al., 2001a,b).

3.4.2. Granite-related Sn and W and hydrothermal Ni deposits of Tasmania (with contributions from R Bottrill)

The diachronous felsic magmatic event in northern and western Tasmania was accompanied by a variety of granite-related mineral deposits. Like the magmatism, which started at ~400 Ma and decreased in age to ~350 Ma to the west, the granite-related deposits (Fig. 60) are older in eastern Tasmania and younger in western Tasmania, with the oldest deposit, the Anchor greisen Sn deposit in northeast Tasmania, and the youngest deposits, the Bold Head and Dolphin W skarns, on King Island. In addition to the above Sn and W deposits, this mineralising event included Sn carbonate replacement deposits (e.g., Renison Bell and Mount Bischoff), magnetite-scheelite skarns (e.g., Kara), Au-Bi skarns (e.g., Stormont), fluorite skarns (e.g., Moina), Sn-W veins deposits (e.g., Aberfoyle and Storeys Creek), and fissure vein Zn-Pb-Ag deposits (e.g., Farrell, Zeehan and Dundas fields). All of these deposits are located within the 4 km granite isobath (Seymour et al., 2006). Global resources from this diachronous event total 0.65 Mt Sn, 0.20 Mt W, 0.10 Mt Zn, 0.18 Mt Pb and 0.15 Mt Ni, as well as small amounts of Cu, Ag, Au, Bi and Mo (data from Seymour et al., 2006 and www.allegiance-mining.com.au [accessed 19 August 2008]). In addition, the Kara skarn deposit has produced significant quantities of high purity iron ore, and the Moina skarn is a major fluorite deposit (18 Mt grading 26% CaF₂, 0.1% Sn and 0.1% WO₃).

The oldest deposits, located in the East Tasmania Terrane, are spatially associated with the ~378 Ma Lottah (Anchor greisen) and the ~377 Ma Henbury (Aberfoyle and Storeys Creek Sn±W veins) granites (Black et al., 2005). The Anchor deposit (2.39 Mt grading 0.28% Sn; Seymour et al., 2006) is hosted by a sheet-like body of muscovite-biotite granite (Lottah Granite) that has intruded porphyritic biotite adamellite of the ~387 Ma (Black et al., 2005) Poimena Granite. The ores are associated with sheet-like greisen lenses that are parallel to, and within 40 m, of the upper contact of the muscovite-biotite granite. The greisen bodies consist of muscovite±topaz±fluorite±siderite altered granite with disseminated cassiterite and variable, but low, amounts of chalcopyrite, bornite, molybdenite and fluorite (Groves and Taylor, 1973). The Aberfoyle vein system (2.1 Mt grading 0.91% Sn and 0.28% WO₃) consists of sheeted quartz veins developed above, or tangential, to the cupola of a greisenised aplite. These veins are hosted by turbiditic rocks of the Mathinna Formation and cut Tabberabberan folds. The quartz veins contain cassiterite, wolframite, fluorite, muscovite, siderite, triplite, sphalerite, chalcopyrite and pyrite with minor stannite and scheelite (Green, 1990; Edwards and Lyon, 1957).

Further to the west, in the eastern West Tasmania Terrane, fluorite (Moina), Zn-Au-Bi (Hugo: 0.25 Mt @ 5-6% Zn, 1 ppm Au and 0.1% Bi) and Au-Bi (Stormont: 0.135 Mt @3.44 g/t Au and 0.21% Bi) skarn deposits are associated with the buried Dolcoath Granite in the Moina area (Morrison et al., 2003). The Kara skarn deposits have produced significant quantities of high purity magnetite ore and scheelite, and other major magnetite skarns occur including the Tenth Legion, St Dizier, Granville, Mt Ramsay and Mt Lindsay skarns, usually containing moderate Sn and/or W grades; these mostly form as magnesian skarns in late Proterozoic dolomites in the Oonah Formation and fringe the Heemskirk and Meredith granites. The distinction between these skarns and the Savage River style magnetite deposits in the Arthur Metamorphic Complex can be blurred.

The most significant granite-related deposits are the Renison Bell (24.54 Mt grading 1.41% Sn) and Mount Bischoff (10.54 Mt grading 1.1% Sn) carbonate replacement deposits (Seymour et al., 2006). These and other smaller deposits are hosted by carbonate units within the Neoproterozoic Oonah Formation and the Neoproterozoic to Cambrian Success Creek and Crimson Creek Formations. Although generally regarded as magmatic-related, these deposits are generally removed from the inferred progenitor granite. At the Renison Bell deposit, a direct link cannot be made with the inferred intrusion, however, at Mount Bischoff, greisenised quartz-porphyry dykes acted as fluid conduits into the dolomitic beds that were replaced by stanniferous massive pyrrhotite (Halley and Walshe, 1995). At Renison Bell, quartz-arsenopyrite-pyrrhotite-cassiterite-fluorite veins infill the Federal-Bassett fault zone, which has acted as a feeder to stratabound, stanniferous massive pyrrhotite ores that have replaced three dolomitic lenses within the Crimson Creek Formation. The ores also contain variable amounts of siderite, talc (in stratabound ores), tourmaline, stannite, pyrite, chalcopyrite, ilmenite, bismuth, fluorite, apatite, sphalerite and galena

(Patterson et al., 1981; Morland, 1990). In contrast, at Mount Bischoff, variably greisenised (extremely topaz-tourmaline altered) quartz-porphyry dykes acted as fluid conduits into the dolomitic beds that were replaced by stanniferous massive pyrrhotite plus zones with variable chondrodite, serpentine, sellaite and talc skarns (Halley and Walshe, 1995). Late veins occur in both deposits, mostly with galena, sphalerite, quartz, fluorite and Mg-Fe carbonates.

Around the Heemskirk, Dolcoath and Meridith granites, carbonate-replacement Sn deposits for part of zoned mineral districts that overlie the Sn-W-mineralised granites. In the Zeehan district, Zn-Pb-Ag vein deposits are zoned around the Queen Hill-Montana-Severn (3.6 Mt grading 1.2% Sn) carbonate replacement deposit (Both and Williams, 1968; Green, 1990). These deposits are most likely related to the emplacement of granite similar in age to the nearby Heemskirk Granite, which has an age of ~361 Ma (Black et al., 2005).

An unusual aspect of the Zeehan district is the presence of a number unusual Ni-only deposits such as Avebury (14.0 Mt grading 1.04% Ni). These deposits consist of disseminated to locally massive sulphides near the contacts of Cambrian ultramafics (of the Trial Harbour Ultramafic Complex) which possibly intruded dolomitic, siliciclastic and mafic rich, conglomeratic sediments, perhaps of the Crimson Creek Formation. The rocks are hornfelsed and altered to skarn-like, amphibolitic and locally pyroxenitic and olivine-bearing rocks. The gangue minerals include serpentine minerals (mostly lizardite and antigorite), talc, chlorite, olivine, pyroxenes, prehnite, amphiboles, axinite, epidote-clinozoisite, sphene, tourmaline, quartz, feldspars, micas (phlogopite and muscovite?), fluorapophyllite, fluorapatite, calcite, dolomite, rutile and spinels (magnetite and chromite-magnesiochromite) (McKeown, 1998, R Bottrill, unpub. data). Veins formed during or following this metasomatic event contain prehnite, axinite, calcite, dolomite, fluorapophyllite, pyrite and other low-T minerals.

The sulphide ore minerals occur in both the serpentinites and amphibolitic rocks. Pentlandite is the most abundant sulphide, unlike most other nickel deposits, which usually dominated by iron sulphides. This makes the deposits metallurgically attractive. Other sulphides in the deposits include minor pyrite, pyrrhotite, chalcopyrite, millerite, mackinawite, sphalerite, a valleriite-haapaite series mineral, minor arsenides (nickeline, gersdorffite and maucherite), and trace native bismuth (McKeown, 1998, R Bottrill, unpub. data). There is also trace cobalt in the ores, probably mostly within the nickel sulphides. Some deposits (e.g., Burbank) contain minor Zn and other base metals. Ores in some deposits in the area (eg. Lord Brassey and Trial Harbour) are dominantly heazlewoodite with minor awaruite, millerite and pentlandite.

The mineralisation probably formed as a result of remobilisation and sulphidation of low grade nickel originally in the ultramafics (possibly as disseminated heazlewoodite and awaruite, as in the Lord Brassey (Heazlewood) and Trial Harbour deposits, and/or from Ni-bearing serpentines and other silicates) into anticlines and other structural traps, mostly on the upper serpentinite contacts. This remobilisation is probably due to the granite-derived hydrothermal fluids, possibly from the adjacent ~361 Ma Heemskirk Granite at Avebury, or the similar aged Meredith granite at Heazlewood. The fluids also altered and hornfelsed the host rocks, and may have introduced some sulphur, arsenic and base metals into the rocks (McKeown, 1998; R Bottrill, unpub. data; Hoatson et al., 2005; pers. comm., 2008). Alternatively, Seymour et al. (2006) suggested that some of the deposits could be termed Ni skarns. However, the protoliths appear to have been largely altered mafic rocks rather than carbonates and the designation is probably inappropriate, although the local skarns may have been cogenetic. They are best termed hydrothermal Ni deposits.

The youngest, and westernmost, granite-related deposits in Tasmania are the Bold Head and Dolphin skarn deposits, which in total had resources of 23.8 Mt grading 0.66% WO₃ (Seymour et al., 2006). These deposits are associated with the Grassy Granite, which has an age of ~351 Ma (Black et al., 2005) and hosted by dominantly siliciclastic rocks of Late Neoproterozoic age that correlate with the Rocky Cape Group (Seymour et al., 2006). In detail, the ores are hosted by skarn and hornfels containing variable

amounts of pyroxene, andradite, grossularite, biotite and calcite. Scheelite, the major ore mineral, is associated with andradite, and pyrrhotite is the main sulphide mineral (Danielson, 1975).

The last major group of deposits associated with Devonian magmatism in Tasmania are fissure vein base metal deposits, which are found in most of the pre-Devonian sedimentary and volcanic sequences of Tasmania and are usually associated with Devonian magmatism. These deposits include the Farrell group of deposits (0.908 Mt grading 2.5% Zn, 12.5% Pb and 408 g/t Ag) and the Magnet deposit (0.63 Mt grading 7.3% Zn, 7.3% Pb and 427 g/t Ag). The Farrell deposits are sheared pyrite-sphalerite-galena-chalcopryrite-quartz lodes hosted by Cambrian slates. The Magnet deposit is a laminated dolomite-siderite-galena-sphalerite vein hosted by mafic volcanic rocks probably related to the Heazlewood Ultramafic Complex. In both deposits, the ores contain moderate amounts of Ag-bearing sulphosalt minerals (Burton, 1975; Cox, 1975). The Zeehan-Dundas mineral fields contains a large number of Magnet style Ag-Pb-Zn rich carbonate veins, the largest possibly being the Comstock deposits (0.1Mt). Host rocks include Proterozoic metasediments through Cambrian ultramafics to Silurian mudstones. There are also some old mines which worked Ag-Pb-Zn veins in Mathinna Beds in the Scamander area, northeast Tasmania, part of a well-zoned Sn-W-Cu-Pb-Zn mineral field (Groves, 1972) related to granites related to the ~400 Ma Blue Tier Batholith.

There are also some notable copper veins which may be granite related, particularly the Balfour deposits (Murrays Reward has a resource of ~0.5Mt @ 0.8% Cu). The Temma deposits are enigmatic magnetite-Fe-Mn carbonate lodes locally rich in Cu, Pb and Zn. These deposits are all hosted in the Proterozoic Rocky Cape Group metasediments and are assumed to be Devonian in age, possibly related to the Sn-W mineralised Interview granite (Taheri and Bottrill, 2005).

3.4.3. Mineral potential

Synthesis and analysis of geological and metallogenic data suggest that the Kanimblan cycle was characterised by two major geodynamic systems:

1. Connors-Auburn arc-backarc system (Fig. 65); and
2. Kanimblan Orogeny (Fig. 66).

Based on existing metallogenic data and the tectonic evolution model presented in section 2.5, predictions are made below about the existence and likely extent of mineral systems for both of these tectonic systems.

3.4.3.1. Connors-Auburn arc-backarc system

Figure 65 shows the mineral potential for the Connors-Auburn arc and related backarc and forearc systems based on the interpretation that the Kanimblan cycle was dominated by west dipping subduction offshore that extended along the eastern margin of Australia at that time (Fig. 40). In this model, the arc is represented by I-type granites that extend inland from the Queensland coast from northwest of Brisbane to just south of Townsville. Basins, including the Yarrol and Tamworth Terranes, that are located outboard of the magmatic arc, are interpreted as forearc basins, and blocks further outboard (e.g., Woolomin, Wisemans Arm, Central and Coffs Harbour blocks in the southern New England Orogen and the Coastal, Yarraman, North D'Aguilar, South D'Aguilar and Beenleigh blocks in the northern New England Orogen) are interpreted as accretionary wedges (Fig. 38). Inboard of the magmatic arc, a backarc environment is interpreted for the early history of Drummond Basin to the north and sedimentary rocks of the Kanimblan cycle in the Lachlan Orogen.

This tectonic framework predicts the presence of porphyry Cu, epithermal and related intrusion-related systems along the Connors-Auburn arc, which extends along the western margin of the New England Orogen (Fig. 65). However, the preservation of these deposits is governed by the depth of exhumation of the host rocks. No significant deposits of these types are known associated with the Connors-Auburn arc, suggesting that these deposits, if formed, may have been stripped by erosion, particularly in the southern

New England Orogen. Known deposits and greater potential for Kanimblan cycle deposits are present inboard from the arc, particularly in the Lachlan Orogen.

Supracrustal rocks deposited during the Kanimblan cycle in the Lachlan Orogen include an early phase of bimodal volcanic and associated rocks, followed by a sequence comprised of clastic, mostly continental, sedimentary rocks, including red beds of the 'Lambie facies' (section 1.1.5). These rocks were deposited as a consequence of continental extensions, possibly related to backarc rifting associated with the Connors-Auburn arc. The bimodal volcanic rocks have potential for epithermal deposits, which is supported by the presence of pyrophyllite-bearing alteration assemblages at the Back Creek mine in the Boyd Volcanic Complex (Lewis et al, 1994). The younger largely continental sedimentary sequence has potential for sediment-hosted Cu and, possibly, U deposits. Vandenberg et al. (2000) documented "many small" $\text{Cu} \pm \text{U} \pm \text{V} \pm \text{Ag}$ occurrences in the Mansfield Basin in eastern Victoria. These occurrences are hosted both by sandstone and shale. Lithologically similar rocks are also present in New South Wales (Merrimbula Group: Lewis et al., 1995), but do not have significant known occurrences of Cu and related minerals.

The most significant known deposits potentially associated with the Connors-Auburn system are located in the West Lachlan, where world class Sn and W deposits are spatially and temporally associated with 380-350 Ma granites in Tasmania (section 3.4.2) and where vein and disseminated gold deposits spatially and temporally overlap magmatic rocks including mineralised dikes where the age of dike emplacement and mineralisation is commonly coeval (section 3.4.1). In Tasmania, the emplacement ages of the granites young to the west (Black et al., 2005), perhaps consistent with a shallowing in the angle of subduction with time. However, in Victoria the granites young towards the Melbourne Zone; the cause of this pattern is not understood (section 1.1.5). Kanimblan-aged gold deposits in Victoria have a close spatial, and in some cases a demonstrated temporal association with granites and temporally these deposits pre-date the Kanimblan Orogeny. Accordingly we consider these deposits intrusion-related and not lode gold deposits and infer that the distribution of Kanimblan-aged granites has a strong control on the location of these deposits.

Although present distribution patterns suggest that Kanimblan cycle intrusion-related deposits in Tasmania are dominated by Sn and W, and similar aged deposits in Victoria are dominated by Au, we consider that some potential for Kanimblan cycle Sn and W deposits exists in Victoria and potential for intrusion-related Au deposits exists in Tasmania. The latter inference is supported by the presence of Au-Bi skarns associated with the Dolcoath Granite (Morrison et al., 2003) and older (Taberraberran Cycle) inferred IRG deposits at Lisle (T Callaghan in Taheri et al., 2003).

3.4.3.2. Kanimblan Orogeny

Although extensively developed through the Tasman Orogen as low-grade metamorphism and east-west shortening, the 360-340 Ma Kanimblan Orogeny is best developed in the East Lachlan, with folding and inversion of Kanimblan cycle basins. Hence, the East Lachlan has the greatest potential for lode Au deposits associated with Kanimblan deformation, although low potential probably exists throughout the Tasman Orogen. The potential for lode Au deposits in the East Lachlan is supported by the presence of 360-345 Ma deposits in the Hill End Trough (section 3.3.8).

In central and north Queensland, the Kanimblan Orogeny does not have a major effect on the Connors-Auburn arc-backarc system, with magmatism and sedimentation largely unaffected by this orogeny. The most significant effect was termination of sedimentation in the Drummond Basin (section 2.6).

3.5. Hunter-Bowen cycle (350 Ma-230 Ma)

Although relatively weakly mineralised in comparison with other cycles, the Hunter-Bowen cycle contains a very diverse assemblage of mineral deposits, many of which are granite-related. Spatially, and to a certain degree temporally, most deposits of the Hunter-Bowen Cycle can be split into two groups: (1) widespread, but mostly small, deposits in northern Queensland associated with Permo-Carboniferous (345-260 Ma)

magmatism of the Kennedy magmatic province, and (2) deposits located mostly in the New England Orogen, with ages mostly between 290 and 230 Ma. In the New England Orogen the deposits appear to be concentrated in two periods: 290-275 Ma and 255-230 Ma. The first period includes ~290 Ma Au-rich epithermal deposits and 280-275 Ma VHMS deposits, whereas the second period is characterised by a more diverse deposit assemblage, including ~255 Ma lode Au-Sb and epithermal Ag deposits, and 250-235 Ma intrusion-related and epithermal deposits. Five hundred km to the west of the New England Orogen, Sn-W and Auebury-type Ni deposits are associated with 235-230 Ma granites in the Doradilla district. The youngest Hunter-Bowen cycle deposits in the New England Orogen are lode Au deposits with a poorly defined Late Triassic (215 Ma) age that slightly post-dates the Hunter-Bowen cycle.

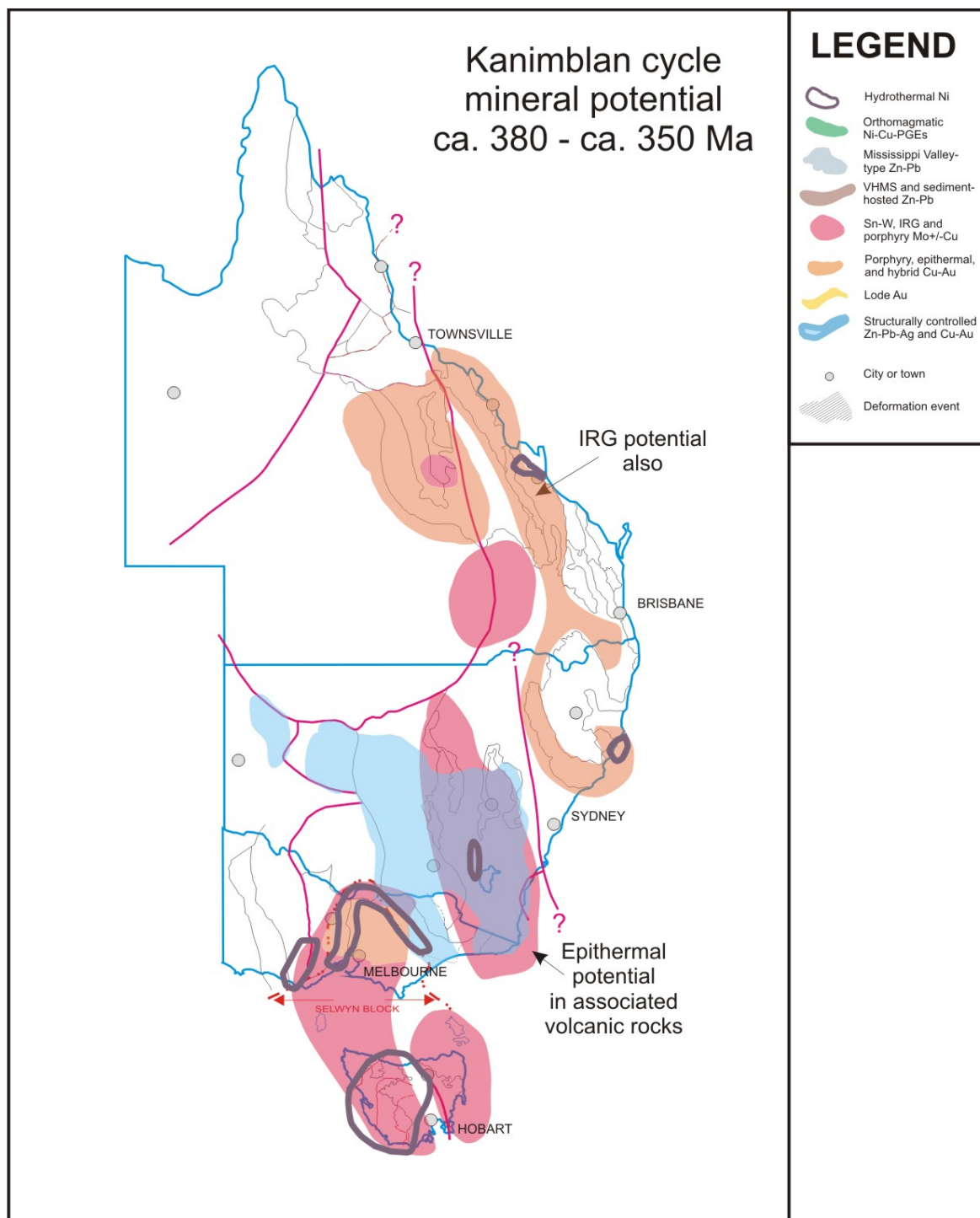


Figure 65. Mineral potential of the Kanimblan cycle.

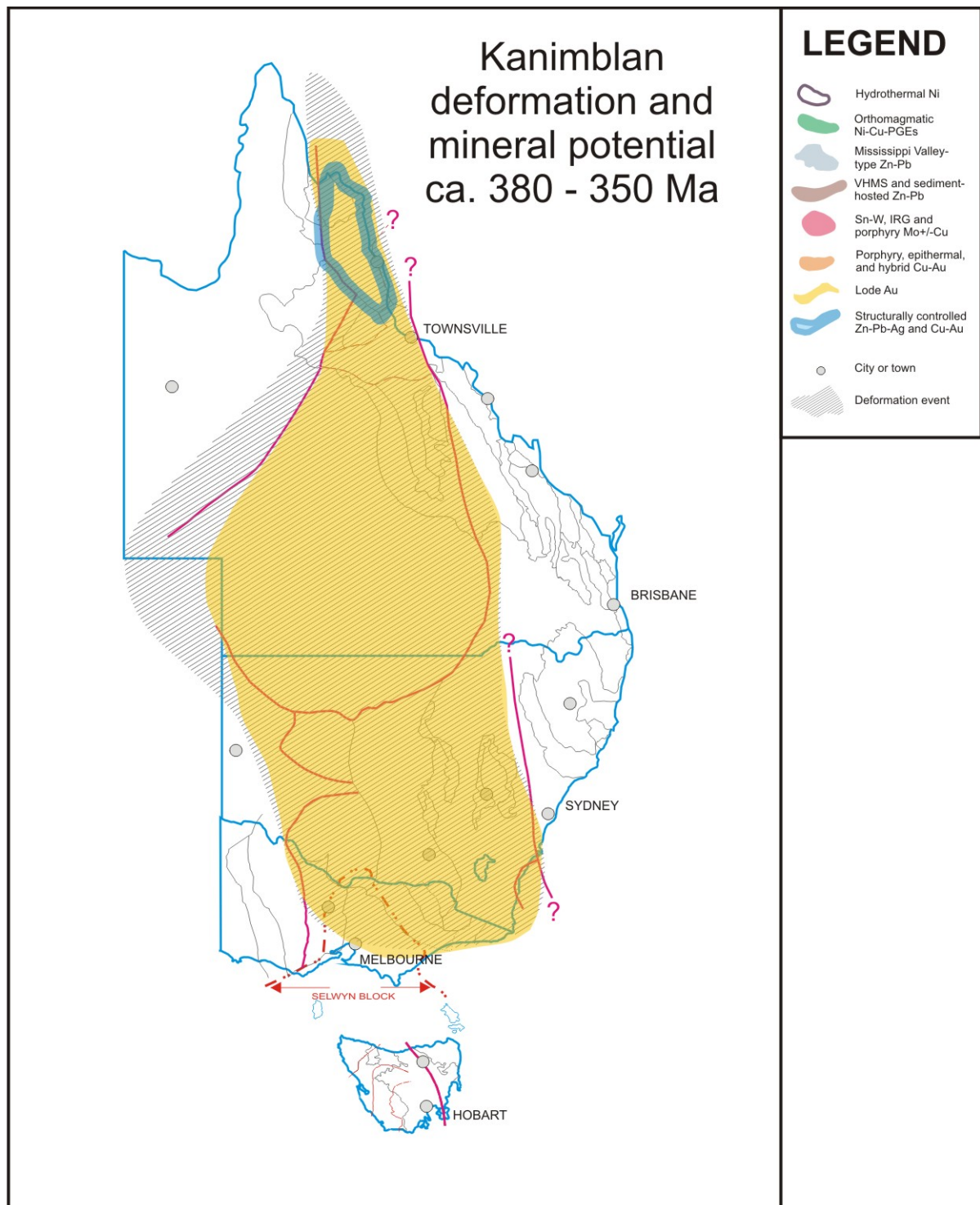


Figure 66. Mineral potential of the Kanimblan Orogeny.

3.5.1. Permo-Carboniferous (345-280 Ma) intrusion-related deposits of north Queensland

Widespread mineralisation occurs associated with Kennedy Province felsic (mostly intrusive) magmatism in northern Queensland (e.g., Gregory et al., 1980; Murray, 1990; Morrison and Beams, 1995). The magmatism-related mineralisation encompasses a variety of styles (Fig. 67), including Sn-W, intrusion-related gold ($\pm$ Mo, Bi, W), porphyry Cu-Au and Cu-Mo $\pm$ base metals, as well as epithermal Au $\pm$ Ag, and perhaps U-F-Mo mineralisation. Although much of the mineralisation is of small size, significant deposits exist for most styles. Much research over the last 25 years has focussed on both the magmatism (e.g.,

Sheraton and Labonne, 1978; Richards, 1980; Champion and Chappell, 1992; Champion and Heinemann, 1994) and the related mineralisation (e.g., Morrison and Beams, 1995; Pollard and Taylor, 1983; Pollard et al., 1983; Witt, 1987, 1988; Blake, 1972). What is evident from the region is that deposit styles although varied, largely relate to the same tectonic and magmatic events, with mineralisation styles and commodity types more related to magmatic controls such as granite composition, oxidation state, and degree of fractionation, as pointed out by Blevin and co-workers (e.g., Blevin and Chappell, 1992, 1996; Blevin, 2004, 2005; Champion and Blevin, 2005), and also by Sillitoe (1991), Thompson et al. (1999), Lang et al. (2000), Lang and Baker (2001). In addition to magmatic-related deposits there are also lode gold deposits such as the Hodgkinson gold field.

3.5.1.1. Sn±W and W±Sn deposits

Extensive and widespread Sn-W mineralisation occurs throughout north Queensland. Most production (and current reserves) has been from four main regions (Fig. 67): the Herberton-Mount Garnet-Georgetown region (Herberton province of Solomon and Groves (2000): Mount Garnet, Herberton-Irvinebank, Sunnymount/Tommy Burns, Koorboora); the Cairns-Cooktown region (Collingwood Province of Solomon and Groves, 2000): Collingwood, Kings Plain); the Mount Carbine-Cannibal Creek region (Mount Carbine province of Solomon and Groves, 2000): Mount Carbine); and the Kangaroo Hills region (Kangaroo Hills Province of Solomon and Groves, 2000): Sardine). Production figures for these regions are >150 kt Sn (~200 kt of tin concentrate; Morrison and Beams, 1995) and 4 kt W (Solomon and Groves, 2000). The vast majority of this has been derived from Herberton province (140 kt of tin concentrate; Murray, 1990) and, in particular, the Herberton-Mount Garnet area (the eastern half of the Herberton Province), with recorded production (up to 1969) of ~110 kt of tin concentrate (Blake, 1972). Although extensive, total Sn production from the region is relatively minor when compared with the larger Devonian Sn deposits of Tasmania (see Solomon and Groves, 2000). Most of the W production in the region has come from the Mount Carbine deposit (Mount Carbine Province), though a number of relatively small W-Mo-Bi pipes (Bamford Hill and Wolfram Camp) have also been mined.

Much of the Sn mined in North Queensland (as cassiterite) has been derived from alluvial and eluvial deposits, including deep leads, over the last 100+ years, within all regions (Withnall et al., 1997a, Bultitude et al., 1997). Significant production from lode deposits only occurred in the Herberton Province.

3.5.1.1.1. Herberton. Blake (1972), Taylor (1979), Pollard and Taylor (1983), Kwak and Askins (1981), Brown et al. (1984), amongst others, have documented a variety of Sn-W mineralisation styles in the Herberton Province (Fig. 67), including:

- Chlorite lodes – the most abundant type, occurring within granite or in host rocks. Assemblages include chlorite-quartz-cassiterite-sericite-sulphides±fluorite-topaz-garnet-tourmaline.
- Greisen – either massive with disseminated cassiterite or as greisen lodes, with quartz-mica-cassiterite-fluorite ± wolframite-monazite-topaz-kaolinite-chlorite-sulphides
- quartz-tourmaline lodes – mostly within host metasedimentary rocks
- complex sulphide lodes – in both granite and host rocks; quartz-cassiterite-pyrite-chalcopyrite ± stannite-chlorite-fluorite-epidote-calcite-carbonate
- fluorite-magnetite-garnet ‘wrigglite’ (F-Sn-W) skarns ± cassiterite – scheelite (e.g., the Gillian Prospect near Mount Garnet). Much of the Sn appears to be within silicate phases such as garnet (Kwak and Askins, 1981; Brown et al., 1984).
- quartz lodes - minor
- disseminated cassiterite or wolframite deposits in granites and pegmatites – minor

As is typical of Sn-W systems, most of the mineralisation is either hosted within granite or close to the granite contact within the local host rocks, such as the metasediments (and locally carbonates) of the Hodgkinson Province. Blake (1972) showed that the Sn deposits of the Herberton region exhibited well-defined district zoning, from an innermost W zone (within the granite), to Sn (granite and host rocks), to Cu, and an outer Pb zone (both within the host rocks). Significant production came from the Herberton Sn fields, and, unlike other Sn producing regions in north Queensland, much of this production was from lode

deposits, although much of this was prior to 1930 (see Pollard and Taylor, 1983). Up to 1972, ~70 kt of a total ~110 kt Sn produced was from lode deposits, mostly from numerous small deposits; the Vulcan Mine being the largest producer with ~14 kt (Blake, 1972). Sizeable reserves (greisen, stockwork and skarns) still exist within the region in a number of prospects (e.g., Sailor, Baalgammon, Gillian: Morrison and Beams, 1995; Garrad et al., 2000).

The Sn±W mineralisation in the region is clearly related to granites (e.g., Pollard and Taylor, 1986; Witt, 1988), in particular to the granites of the various suites of the O'Briens Creek Supersuite (Champion and Chappell, 1992; Champion and Blevin, 2005). It is also clear that there is a strong host rock control. Granites of the O'Briens Creek Supersuite are widely distributed (e.g., Champion and Heinemann, 1994); Sn-W mineralisation, however, is strongly localised east of the Palmerville Fault, where granites intrude metasediments of the Hodgkinson Province.

3.5.1.1.2. Kangaroo Hills. Total production from this region (Fig. 67) was ~7.5 kt of Sn (mostly cassiterite, minor stannite) concentrates (Murray, 1990), from alluvial and lode deposits, with the most significant lode deposit being the Sardine Mine (Wyatt et al., 1970). Lode deposits, mostly veins and pipes, include chlorite (dominant), tourmaline and sulphide-rich lodes, greisens, pegmatitic segregations, and skarns (e.g., Wyatt et al., 1970; Gregory et al., 1980; Rienks et al., 2000), within the granites and local sedimentary country rocks. These comprise cassiterite ± base metals but also include stannite (Wyatt et al., 1970). Mineralisation appears to be mostly related to granites of the Oweenee Supersuite, but possibly also to the poorly outcropping S-type Dora Supersuite (Champion and Heinemann, 1994).

3.5.1.1.3. Cooktown. The Cooktown Sn province (Fig. 67), has historically been a minor producer – about 14 kt of largely alluvial Sn concentrate (Murray, 1990). Significant reserves, however, exist, including Collingwood (currently in production – 28 kt Sn) and Jeanie River (54 kt Sn) – both lode deposits and Kings Plain (alluvial - ~6 kt tin concentrate); e.g., Morrison and Beams, (1995), Garrad et al. (2000). Mineralisation at Jeannie River, described by Lord and Fabray (1990), is hosted within the metasediments of the Hodgkinson Formation, and is dominated by complex sulphide lodes, i.e., cassiterite-sulphides (including pyrite-pyrrhotite-sphalerite-galena-chalcopyrite-arsenopyrite). Zoning is evident, from an outer Pb-Zn±Sn zone to an inner pyrrhotite-rich, Cu-As-W±Sn stockwork vein swarm zone (Lord and Fabray, 1990). Alteration includes silicic, sericitic and propylitic assemblages (Lord and Fabray, 1990). Although thought to be intrusion-related, the granites responsible for the mineralisation have not been unequivocally identified. Lord and Fabray (1990) favoured porphyry dykes that occur near some of the prospects.

Mineralisation at Collingwood, south of Cooktown, has been described by Jones et al. (1990). The deposit occurs within an area of Hodgkinson Formation metasediments intruded by S-type granites of the Cooktown Supersuite. Mineralisation is largely hosted within various flat-lying phases of the Finlayson Batholith. Jones et al. (1990) document three mineralisation styles:

- Steep siliceous sheeted greisen veins, and associated siliceous (quartz-muscovite-biotite-tourmaline) alteration haloes, which host most of the ore.
- High-grade albitic veins (albite-cassiterite-chlorite-fluorite± biotite-muscovite-sulphides assemblages) with siliceous alteration haloes that cross-cut the sheeted veins
- Sub-horizontal mineralisation and silica, tourmaline and sericite alteration, along the granite-sediment contact, associated with topographic highs in the granite.

Tin mineralisation increases in grade towards the granite contact but is minor outside of the granites (Jones et al., 1990). Mineralisation in the Cooktown tinfield is related to the Permian S-type granites of the Cooktown Supersuite (Bultitude and Champion, 1992; Champion and Blevin, 2005).

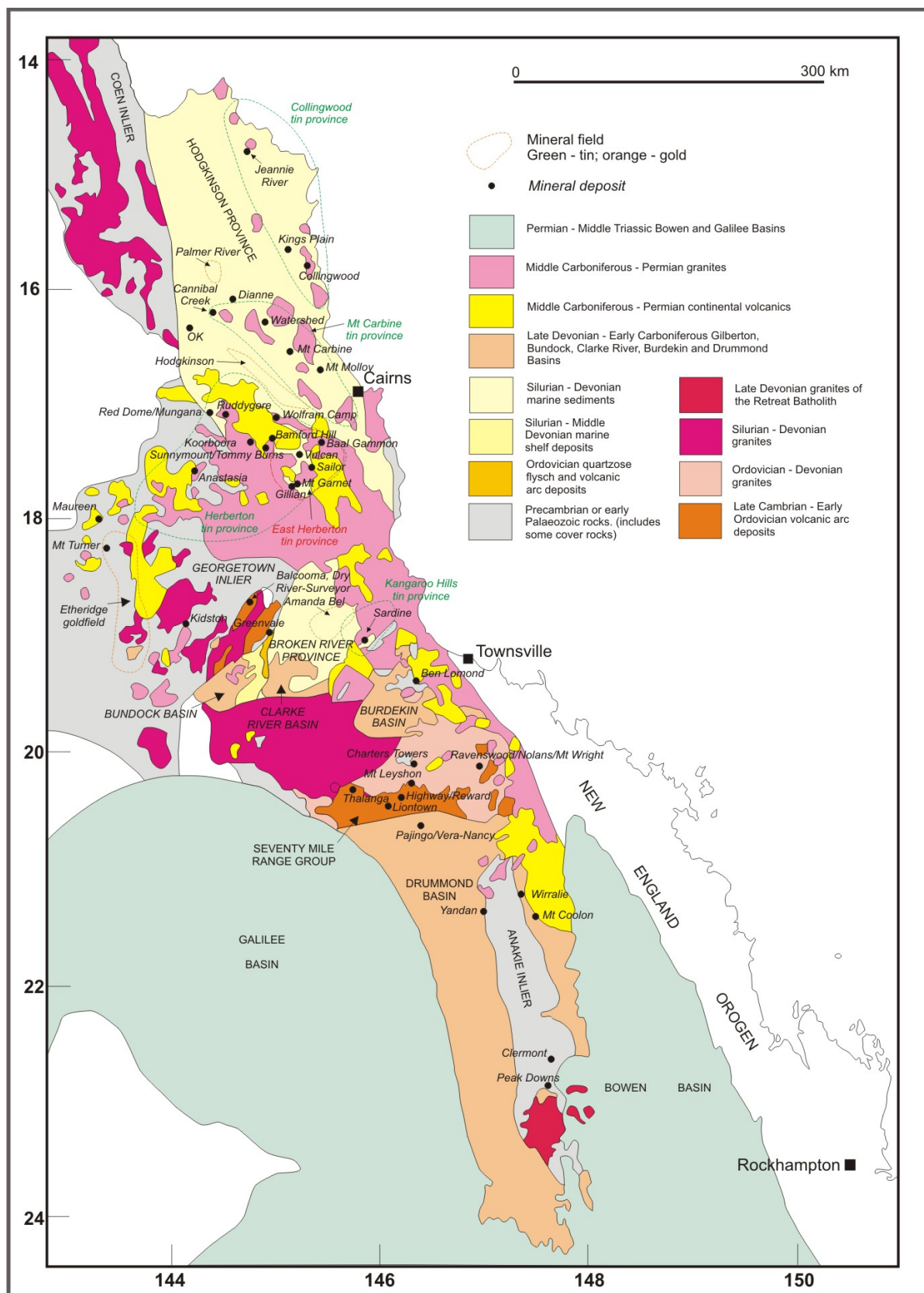


Figure 67. Distribution of Sn-W and Au-Mo-Cu-Bi mineral provinces and major individual deposits associated with Kennedy province magmatism in north Queensland. Modified after Murray (1990).

3.5.1.1.4. Mount Carbine Province. This province (Fig. 67) is characterised by Sn-W and W-Sn mineralisation, with a number of significant W deposits, of which Mount Carbine deposit is the most important. The province is largely delineated by the distribution of the S-type Whypalla Supersuite (Bultitude and Champion, 1992; Champion and Bultitude, 1994) to which mineralisation is spatially and probably genetically related (e.g., Higgins et al., 1987). Total production includes alluvial and lode Sn-W and W mineralisation, mostly from veins and alluvial deposits but also skarns (Solomon and Groves, 2000). The province differs from the others in the presence of significant W mineralisation (e.g., Mount Carbine and Watershed).

Mount Carbine, northwest of Cairns, is a wolframite deposit hosted by the Silurian to Devonian metasediments of the Hodgkinson Formation. As described by Forsythe and Higgins (1990), mineralisation occurs as multiple vein zones comprising early, steep-dipping, quartz-apatite±wolframite±K-feldspar±biotite±muscovite±molbdenite±bismuth veins with strong tourmaline-biotite alteration selvages and later (partly overprinting) fluorite-chlorite-tourmaline-albite-scheelite-cassiterite-sulphide-calcite vein and fracture-fill assemblages. About 16.4 kt of wolframite concentrate was produced from the deposit (Murray, 1990). Mineralisation is thought to be largely genetically related to intrusives, most probably the nearby Permian S-type Whypalla Supersuite granites in the Mossman batholith (Bultitude and Champion, 1992), consistent with ages of ca. 280 Ma for the local granite and greisen alteration (see summary in Forsythe and Higgins, 1990).

The Watershed prospect, northwest of Mount Carbine, contains scheelite mineralisation hosted by the Silurian to Devonian metasediments of the Hodgkinson Formation. Scheelite is present as disseminations within altered metasediments including calc-silicates, and within quartz-feldspar veins (Pertzel, 2007). The deposit has been interpreted as a quartz-vein swarm (Pertzel, 2007). Current resource estimates are ~44.1 kt contained WO₃ (<http://www.vitalmetals.com.au/projects/watershed.phtml> [accessed December, 2008]), although mineralisation is apparently open along strike and at depth.

Other W deposits occur in the region, most notably W-Mo-Bi mineralisation, which occur as pipe-like bodies and associated greisens within granites in the Herberton region (e.g., Bamford Hill and Wolfram Camp (refs)), and in the Kangaroo Hills region (Rienks et al., 2000). Deposits were mostly small: Wolfram Camp and Bamford Hill together, produced ~9.3 kt wolframite concentrates and <2 kt molybdenum concentrates (Murray, 1990). In the Herberton region W-Mo-Bi mineralisation is associated with granites of the Ootann Supersuite (e.g., Blevin, 2004). One feature of these deposits is the large grain size of many of the ore minerals within the pipes.

Tin-tungsten mineralisation in all regions is related to the associated granites (i.e., granites of the O'Briens Creek Supersuite in the Herberton-Mount Garnet-Georgetown region; the Kangaroo Hills region – granites of the Oweenee and Dora Supersuites largely (Rienks et al., 2000) the Cooktown region (including Collingwood) – granites of the Cooktown Supersuite, and the Mount Carbine region (including Mount Carbine) – granites of the Whypalla Supersuite (Blake, 1972, Richards, 1980; Champion and Bultitude, 1994; Champion and Heinemann, 1994)). As shown by Champion and Blevin (2005) there is a clear strong relationship in North Queensland between Sn mineralisation and strongly fractionated, reduced I- and S-type granites.

3.5.1.2. Porphyry Cu-Mo-Au and intrusion-related gold deposits

A range of styles of Carboniferous to Permian intrusion-related gold mineralisation exist in North Queensland. Morrison and Beams (1995) included these all together under a broad 'Porphyry' category, but subdivided them into breccia, stockwork, vein and skarn subtypes. Recent work on intrusion-related gold deposits, chiefly in North America, has led to the recognition of an intrusion-related gold (IRG) mineral system with a wide variety of deposits styles (e.g., Thompson et al., 1999; Lang et al., 2000; Lang and Baker, 2001), that encompass all the north Queensland sub-types. Subsequent work in Australia (e.g., Blevin, 2004, 2005) and by the original proponents of the IRG model, have resulted in identification of two

end-members for intrusion-related gold deposits: “porphyry” Au and Cu-Au deposits associated with primitive oxidised magmas largely in primitive tectonic settings (e.g., Macquarie Arc deposits in NSW), and the IRG end-member with more evolved, more reduced magmas, largely in continental settings. In a north Queensland context the majority of intrusion-related gold deposits belong to the IRG end-member (e.g., Blevin, 2005). The exception is the Mount Leyshon deposit, which is thought to belong to the porphyry Cu-Au porphyry-style (P.Blevin, cited in Champion, 2007). This subdivision is followed here. It should be noted that many IRG deposits in Australia have associated early high temperature molybdenum (Mo-W-Bi) mineralisation with, typically, more distal gold (Blevin, 2005, written commun., 2008).

3.5.1.2.1. Intrusion-related gold (IRG) – Kidston, Red Dome and Mungana. Significant IRG mineralisation occurs within North Queensland associated with the Carboniferous-Permian granites of the Kennedy Province (Fig. 67), and most of the more significant gold producers in the region, such as Kidston, Red Dome and Ravenswood, fall into this category. As with other granite-related mineralisation in the area, specific granite supersuites appear more conducive to this style of mineralisation, in particular the Ootann and closely related Glenmore Supersuites (Champion and Heinemann, 1994; Withnall et al., 1997a), largely reflecting the intensive parameters of the granites in these supersuites (e.g., Blevin et al., 1996). Recorded ages of mineralisation (e.g., Perkins and Kennedy, 1998) range from 330 Ma (Kidston) to ca. 280 Ma (Ravenswood/Mount Wright), consistent with the ages of associated magmatism and Kennedy Province magmatism in general. The mineralisation ages also young to the east and north-east as does the magmatism (see section 1).

The *Kidston* gold deposit, south of Georgetown (Fig. 67), is a breccia pipe deposit thought to be related to Carboniferous magmatism (see Baker and Tullemans, 1990; Baker and Andrew, 1991; Morrison et al., 1996; Bobis et al., 1998). The breccia pipe occurs within the Mesoproterozoic Einasleigh Metamorphics but close to various intrusive phases of the Silurian Oak River Batholith (Withnall and Grimes, 1995), and clasts of these units occur within the breccia. The breccia pipe is of significant size – up to a kilometre-wide at surface, and continuing at depth up to 1.4 km (Bobis et al., 1998). Morrison et al. (1996) and Bobis et al. (1998) have shown that the breccia pipe is zoned from carbonate-pyrite±pyrrhotite at the top, through a number of zones (including gold-base metals), to quartz-fluorite-molybdenite-Bi phases-pyrrhotite±base metals±scheelite±wolframite at depth. The deposit becomes more porphyry-Mo like at depth (Bobis et al., 1998). Gold apparently overprints Mo mineralisation, but both are suggested to be intrusion-related (e.g., Morrison et al., 1996; Bobis et al., 1998). As indicated by Baker and Tullemans (1990), both the brecciation and mineralisation are temporally and spatially associated with the intrusives. Perkins and Kennedy (1998) suggest ca. 332 Ma ages for mineralisation and post-ore dykes at Kidston confirming this relationship. Mineralisation, therefore, is thought to be related to the Carboniferous intrusive phases which occur within the breccia and outcrop nearby (e.g., Withnall and Grimes, 1995; Baker and Tullemans, 1990). The Carboniferous intrusives range from mafic to felsic, and appear to form a largely co-magmatic fractionated suite, though with some isotopic evidence of open-system behaviour (Blevin, 2005) – originally placed in the Ootann Supersuite by Champion and Heinemann (1994) and Blevin (2005), but reassigned by Withnall et al. (1997a) to the Glenmore Supersuite.

The *Red Dome* gold deposit, northwest of Chillagoe (Fig. 67), is a skarn deposit related to intrusion of Carboniferous rhyolite dykes, at high crustal levels (e.g., Ewers et al., 1990; Nethery and Barr, 1998), into steeply-dipping limestone and other sediments of the Chillagoe Formation, Hodgkinson Province (Bultitude et al., 1997). The deposit falls at the eastern end of a west-northwest trending belt of deposits – the Mungana Group (Nethery and Barr, 1998). Within the deposit two separate episodes of rhyolite intrusion and associated magnetite-bearing skarn development have been documented (Ewers et al., 1990), as well as an intermediate phase of wollastonite-bearing skarn and retrogressive alteration after each skarn phase (Ewers and Sun, 1988; Ewers et al., 1990). Nethery and Barr (1998) also indicate a late epithermal phase perhaps related to A-type intrusions. The deposit also comprises an upper post-mineralisation breccia zone, thought to be a collapse breccia (Ewers et al., 1990). Recorded sulphide minerals include bornite-chalcocite (with wollastonite) and chalcopyrite-arsenopyrite-pyrite-sphalerite elsewhere (Ewers et al., 1990). Gold appears to be largely associated with the wollastonite-bearing skarn and first phase of post-skarn-retrogression (Ewers et al., 1990; Nethery and Barr, 1998), as well as in quartz-arsenopyrite-

molybdenite±gold stockwork mineralisation spatially associated with the intrusives. The second rhyolite phase appears not to have introduced additional gold but only remobilised existing mineralisation (Ewers et al., 1990) though Nethery and Barr (1998) appear to indicate Au introduction (and base metals) with the second retrogressive phase also. Fluid inclusion data indicate temperatures to 380°C and the presence of high salinity fluids (Ewers and Sun, 1989). The Red Dome mine produced over 30 t Au and 70000 t Cu, and has a current resource of ~8.5 Mt at 1.61g/t Au, 0.4% Cu and 13 g/t Ag (<http://www.kagara.com.au> [accessed December 2008]). Kagara Zinc report the presence of molybdenum mineralisation at depth. Nethery and Barr (1998), largely on the basis of the widespread nature of intrusive-related alteration assemblages in the region, suggested that the porphyries at Red Dome, and in the Mungana group deposits and environs in general, were related to a larger pluton(s) at depth. These authors indicate similar mineralisation to Red Dome also occurred within other deposits in the region, such as Mungana. As pointed out by Nethery and Barr (1998), however, the latter also appears to contain earlier Besshi-style VHMS mineralisation, which have been overprinted by the later Carboniferous-Permian intrusive activity, skarn development and subsequent retrogression. Like Red Dome, quartz-arsenopyrite-molybdenite±gold mineralisation occurs spatially associated with intrusives, as well as younger epithermal overprinting and acid leaching (Nethery and Barr, 1998). The combined current inferred Au and Ag resources at Red Dome and Mungana are 43 t Au, 426 t Ag and 79 kt Cu (<http://www.kagara.com.au>). Available geochronology (Perkins and Kennedy, 1998; Georgees, 2007) suggests intrusives are ca. 320 Ma in age, though chemical and field considerations suggest a range of intrusive ages from ca. 317 Ma to 307 Ma, and perhaps younger, to ca. 290 Ma (see Nethery and Barr, 1998). Alteration at Mungana and Red Dome, lies around ca. 310 Ma (Perkins and Kennedy, 1998). Georgees (2007) reports a Re-Os molybdenum age of ca. 307 Ma for Mungana.

The *Ravenswood area* (Fig. 67), south of Townsville, has had a long history of gold mining, with some 28 t Au produced from the area prior to 1968 (Collett et al., 1998), from alluvial and lode deposits. Lode gold production was mainly from quartz-sulphide (pyrite-sphalerite-chalcopyrite-arsenopyrite-pyrrhotite) veins, often with high grade gold (up to 30-100 g/t), and minor but intense sericite, calcite and chlorite alteration selvages (McIntosh et al., 1995). Collett et al. (1998) suggested these veins record multiple events with up to eight alteration events and brecciation recognised. Locally the veins form stockworks such as at the Nolans Deposit (McIntosh et al., 1995; Collett et al., 1998). Additional gold has been derived from older, low-grade, chlorite-silica shear zone hosted, locally brecciated, quartz (buck) reefs with pyrite-sphalerite-pyrrhotite-chalcopyrite-gold and a wide biotite/chlorite alteration halo (McIntosh et al., 1995; Collett et al., 1998). All mineralisation is hosted within older Devonian mafic and felsic intrusives of the Ravenswood Batholith. Renewed production since 1987 (11.2 t Au produced for the period 1987-1996; Collett et al., 1998) has been from extensions of old deposits and a number of new deposits such as Nolans (5.1 t Au). Significant reserves (17.85 t Au) and resources (inferred 33.6 t Au) exist in the Nolans and Sarfield deposits (Collett et al., 1998).

The *Mount Wright* deposit, ~10km from Ravenswood, comprises a brecciated Carboniferous-Permian rhyolitic intrusive, which occurs as a pipe-like deposit within older intrusives of the Ravenswood Batholith. Both the rhyolite and host granites have been brecciated (Harvey, 1998). Both brecciation and gold mineralisation, which has a significant vertical extent (up to a kilometre at depth (A-Izzeddin et al., 1995)), appear to be closely associated and related to the rhyolitic intrusives (Harvey, 1998). Mineralisation occurs dominantly as quartz-carbonate (siderite)-sericite-sulphide (arsenopyrite-pyrrhotite-sphalerite-chalcopyrite-molybdenite)-gold breccia infill and minor veins (A-Izzeddin et al., 1995; Harvey, 1998). Alteration assemblages, which extend into the older granite wall rocks, include sericite, silica, chlorite and carbonate (siderite). Historic production, centred on the mother lode, total some 0.5 t Au (Harvey, 1998). The inferred resource (at 1998) was 10 Mt at 3 g/t Au (30 t Au: Harvey, 1998). Total production figures are uncertain, although over 8 t Au has been produced from Ravenswood and Mount Wright (combined) in the last 2 years, with reserves of 5.6 t Au and indicated/inferred resources of over 21 t Au (<http://www.resolute-ltd.com.au> [accessed December 2008]). Geochronology by Perkins and Kennedy (1998) suggests vein mineralisation at Ravenswood and breccia mineralisation at Mount Wright was largely coeval, with ages of ca. 310-305 Ma. Buck reef mineralisation appears to have been earlier – ca. 330 Ma (Perkins and Kennedy, 1998).

3.5.1.2.2. Porphyry style Cu±Mo±Au and Mo±Cu±Au deposits. Despite the great abundance of granites in north Queensland, porphyry-style mineralisation only occurs sporadically (e.g., Horton, 1982). This is particularly true for Cu-rich porphyries. Where mineralisation is known (e.g., Mount Turner, Ruddygore) alteration is often well developed, extensive, and zoned (like porphyry Cu deposits elsewhere), but does not appear to have significant metal endowment (Richards, 1980; Baker and Horton, 1982). The lack of significant Carboniferous-Permian porphyry Cu mineralisation in north Queensland is not unexpected given the general relationships between intensive granite parameters and mineralisation styles (e.g., Blevin et al., 1996; Lang et al., 2000). In general, the great majority of Carboniferous and Permian granites of northern Queensland are too evolved (too felsic) and not oxidised enough to generate significant porphyry Cu ± Mo deposits (e.g., Champion and Blevin, 2005). Similarly, the interpreted tectonic setting (mature continental and probably distal from arc, especially in the Georgetown region; see [section 2](#)) is not consistent with such deposits (Champion, 2007). It is noteworthy that the most significant porphyry-style Cu-Mo deposit in the region is Mount Leyshon (south of Charters Towers), which was mined for its gold. It should be noted that there is some significant Cu-Zn mineralisation in the region. These occur in skarns, the best example of which is Mount Garnet (Hartley and Williamson, 1995). These skarns, which are distinct from the F-Sn-W skarns and associated with felsic granites (Brown et al., 1984), appear to be typically related to Almaden Supersuite granites (the same supersuite responsible for the Ruddygore porphyry mineralisation). Some of the base metal mineralisation in the Mungana region may also relate to Almaden Supersuite granites.

The nature of the Carboniferous-Permian intrusives in north Queensland does not rule out porphyry Mo mineralisation. Although no significant porphyry Mo deposits are known (e.g., Horton, 1982), it is evident that some of the intrusion-related gold deposits, such as Kidston and possibly Red Dome, become porphyry Mo-like at depth (e.g., Bobis et al., 1998).

The *Mount Leyshon* gold deposit (Fig. 67) represents an intrusion-related, hydrothermal breccia-pipe – the Mount Leyshon Breccia Complex – up to 1.5 km in diameter (e.g., Paull et al., 1990; Orr, 1995; Morrison and Blevin, 2001; Allan et al., 2004). Like Kidston, the breccia pipe is located in older rocks, near the (complex) contact between Cambrian-Ordovician metasediments of the Puddler Creek Formation and the Ordovician Fenian Granite (Hutton and Rienks, 1997; Allan et al., 2004). Numerous Carboniferous-Permian porphyries and dyke swarms as well as four breccia phases have been delineated around and within the deposit (e.g., Paull et al., 1990; Orr, 1995; Morrison and Blevin, 2001; Allan et al., 2004). Alteration at Mount Leyshon is largely chlorite and biotite-magnetite (early propylitic) with later feldspar-phyllic alteration, associated with gold, and apparently related to some of the porphyries (Paull et al., 1990; Orr, 1995; Morrison and Blevin, 2001; Allan et al., 2004). Allan et al. (2004) report fluid inclusion data from early quartz-molybdenite-pyrite-chalcopryrite veins, quartz-K-feldspar-chlorite-carbonate-fluorite-pyrite-sphalerite breccia infill, and subsequent sphalerite-pyrite-quartz-gold veins (ore). Fluids associated with the ore are of low to moderate salinity with temperatures of 350-400°C (Allan et al., 2004). According to Morrison and Blevin (2001) mineralisation extends over some 700m vertical extent, containing over 90 t Au (~70 Mt at 1.43 g/t Au). The deposit is zoned from an outer Zn±Au zone, through a Cu-Pb-Zn±Au zone to a pyrite-rich, base metal-poor core (Morrison and Blevin, 2001). Mineralisation at Mount Leyshon was dated at ca. 290-280 Ma by Perkins and Kennedy (1998), and Paull et al. (1990) report ages of ca. 280 Ma for alteration. Lead model ages (J.A. Dean and G.R. Carr, cited in Paull et al., 1990) also suggest ca. 280 Ma ages. More recently, Mugulov et al. (2008) reported a U-Pb zircon age of ~289 Ma for a dyke intimately associated with gold mineralisation. As pointed out by numerous authors (e.g., Paull et al., 1990) these ages are contemporaneous with associated magmatism in the region, supporting intrusion related gold models for the deposit. Morrison and Blevin (2001) suggest Mount Leyshon is a gold-rich porphyry Cu-Mo system. These authors document a co-magmatic suite of andesitic to rhyolitic intrusive phases that appear to be associated with phyllic alteration and gold mineralisation. These intrusives form part of, and lie along, north-east striking corridors of Carboniferous-Permian intrusive and extrusive magmatism which have been considered important in localising gold mineralisation in the region (e.g., Paull et al., 1990).

3.5.2. Middle Carboniferous (~340 Ma) epithermal gold-silver deposits, north Queensland

Although epithermal deposits occur in various places within North Queensland and the Thomson Orogen (e.g., Anastasia, SW of Chillagoe: Nethery, 1998), they are best developed south of Charters Towers in the northern Drummond Basin in the Thomson Orogen (Fig. 67), where significant deposits occur including Pajingo and Vera-Nancy (Fig. 67), Wirralie, and Yandan. These and other deposits in the basin have been recently summarised by Denaro et al. (2004). All are hosted by Cycle 1 Upper Devonian volcanic rocks and sediments of the Drummond Basin (Denaro et al., 2004), though the hosts for the Pajingo deposits may be Upper Devonian to Lower Carboniferous (Richards et al., 1998). They all comprise low-sulphidation quartz-adularia epithermal vein and/or replacement deposits (Porter, 1990; Richards et al., 1998; Seed and Ruxton, 1998; Ruxton and Seed, 1998). At Pajingo and Vera Nancy (Pajingo epithermal system of Mustard et al., 2003), mineralisation occurs as veins. These comprise chalcedonic or cryptocrystalline quartz, clay (illite-kaolinite), carbonate and pyrite, and contain gold and silver (Porter, 1990; Richards et al., 1998). Veins dip at moderate to steep angles, and mineralisation at Vera-Nancy is recorded to 400+ m depth (Porter, 1990; Richards et al., 1998). Alteration envelopes around veins vary from distal chlorite-dominant propylitic (chlorite-calcite $\pm$ pyrite) to proximal, higher temperature, silica-pyrite-sericite-clay (illite, illite-smectite, local kaolinite) phyllic or argillic assemblages (Porter, 1990; Richards et al., 1998; Mustard et al., 2003). Late stage alteration assemblages include carbonate (ankerite, dolomite, siderite) and overprint earlier assemblages (Porter, 1990; Mustard et al., 2003). Mustard et al. (2003) suggested the propylitic alteration, at least in part, represents an earlier, basin-wide alteration event. A variety of quartz vein textures are present in the deposits but most gold (in the Vera-Nancy system) is apparently associated with early-formed banded quartz veins (Mustard et al., 2003). Available fluid inclusion data – summarised in Richards et al. (1998) – suggest low salinity fluids, with evidence for fluid boiling. Brecciation is also recorded (Richards et al., 1998). Location of veins appears to be structurally controlled, which Richards et al. (1998) interpreted as a strike-slip fault. Sericite (K-Ar) ages for the Scott lode indicate middle Carboniferous ages (342 ± 3 Ma) for the mineralisation (Richards et al., 1998). The Scott and Cindy lodes yielded 12 t Au and 38.9 t Ag (Richards et al., 1998). Reserves, in 1998, for Vera and Nancy were 33.5 t Au and 30.9 t Ag (Richards et al., 1998). Current (2008) reserve and resource estimates for the area are 15.1 t Au (<http://www.nqm.com.au> [accessed December 2008]).

Mineralisation at Wirralie comprises quartz-chalcedony-pyrite $\pm$ breccia veins, and stockworks of such veins, as well as silica replacement zones of volcanoclastic hosts (Fellows and Hammond, 1990; Seed, 1995b; Seed and Ruxton, 1998). Alteration includes illite-pyrite and adularia (associated with gold) and overprinting quartz-kaolinite assemblages (Seed and Ruxton, 1998). Sulphides are largely pyrite but include marcasite, chalcopryrite, arsenopyrite, and sphalerite (Fellows and Hammond, 1990; Seed, 1995b). Mineralisation at Wirralie is inferred to have a strong structural control, particularly by reactivated early graben structures (Seed and Ruxton, 1998). Mining at Wirralie to 2001 produced 14.85 t Au from oxide ores (<http://www.ashburton-minerals.com.au> [accessed December 2008]), although Denaro et al. (2004) indicate that 17.3 t of bullion were produced. Current remaining gold resources (measured, indicated and inferred) at Wirralie (at 2004) are 4.7 t Au in oxide zones and 12.35 t Au in sulphide zones (<http://www.ashburton-minerals.com.au> [accessed December 2008]).

Mineralisation at Yandan comprises silica-adularia-illite-pyrite or kaolin-illite-adularia-quartz deposits, largely replacing calcareous sandstone (Ruxton and Seed, 1998). Subordinate, locally brecciated, chalcedonic and quartz-adularia-calcite veins also host ore. Alteration (for the bulk of the ore) comprises a quartz-adularia core with pyrite $\pm$ chalcopryrite (and gold), through illite-smectite, adularia and kaolinite assemblages to a propylitic celadonite-carbonate-chlorite-clay halo (Seed, 1995a; Ruxton and Seed, 1998). Mineralisation appears to be both structurally and lithologically controlled (Ruxton and Seed, 1998). Mining figures for Yandan are variable. Ashburton Minerals reports production of 11.4 t Au from 1992 to 1998 (<http://www.ashburton-minerals.com.au>), whereas Denaro et al. (2004) report 7.03 t of gold bullion for 4.64 t of Au from the period 1993-2001. Delta Gold Ltd (2000; cited in Denaro et al., 2004) report a measured resource of 0.83 t Au at a grade of 0.4 g/t.

As summarised by Denaro et al. (2004), epithermal mineralisation has a strong structural control, though lithology (e.g., Yandan) can also be important. Denaro et al. (2004) suggest north, north-east and south-east structures are most dominant. Fluid boiling appears to have been the most important mechanism for depositing gold (Richards et al., 1998; Ruxton and Seed, 1998; Seed and Ruxton, 1998; Denaro et al., 2004). Although geochronology is sparse most evidence suggests an early (to mid?) Carboniferous age for mineralisation, during the end of Cycle 1 magmatism in the Drummond Basin (e.g., Denaro, et al., 2004).

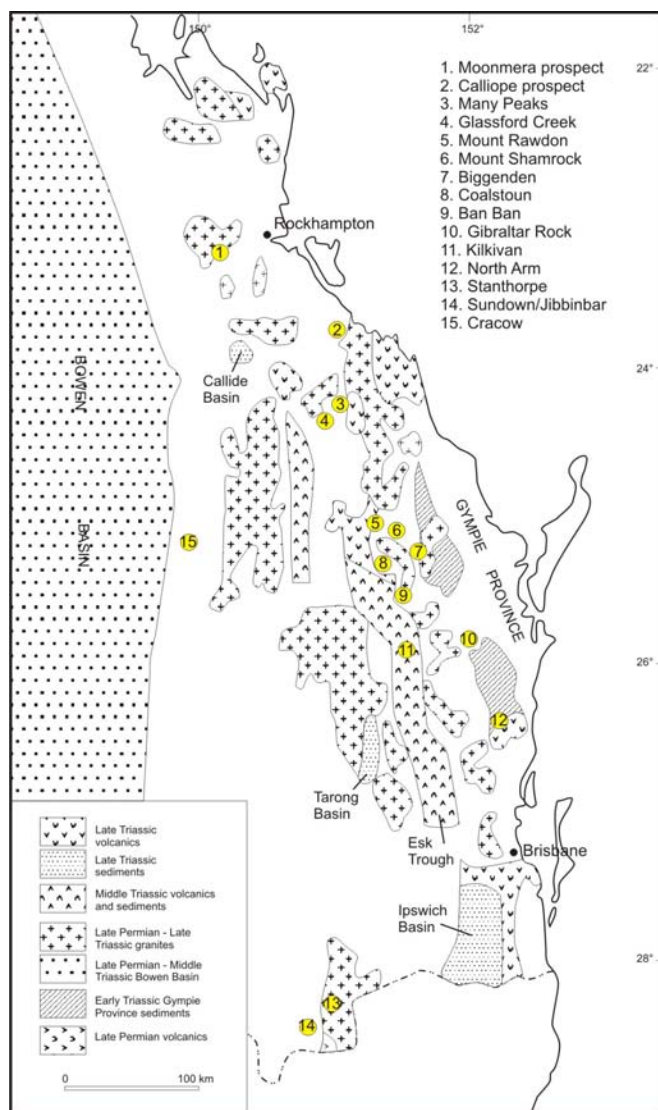


Figure 68. Distribution of Late Permian-Triassic rocks and associated Permian-Triassic mineral deposits discussed in this report for the northern New England Orogen (modified after Murray, 1986).

3.5.3. Early Permian (~290 Ma) epithermal gold deposits, New England Orogen

Late Carboniferous to Early Permian volcanic belts in the New England Orogen contain epithermal auriferous veins at Cracow in south-central Queensland and at Mount Terrible in north-central New South Wales. The Cracow goldfield (Fig. 68), which has produced some 26 t Au, is hosted by gently-dipping felsic and intermediate rocks of the Late Carboniferous-Early Permian Camboon Volcanic Arc (Murray, 1986; 1990). A series of quartz vein systems extend over a NW-trending zone about 5 km long. In general, orebodies occur as open-space vein fillings, which dip vertically to sub-vertically and are structurally controlled (Dong and Zhou, 1996). Most of the Au ore is contained within the east-west Golden Plateau

lode system (1000 m long × 30 m wide) which coincides with a zone of quartz veining, brecciation and silicification (Dong and Zhou, 1996).

The Golden Plateau system is a typical quartz-adularia type epithermal deposit hosted by andesites and volcanoclastics of the Camboon Volcanics (Cracow Mining Venture Staff et al., 1990). Rhyolitic feldspar porphyry and quartz-feldspar porphyry intrusives are spatially associated with the quartz-breccia lodes in the region, and rhyolitic dykes in the eastern part of the Golden Plateau mine appear to be closely related in time to the later, mineralised stages of quartz veining (Cracow Mining Venture Staff et al., 1990). As such, Dong and Zhou (1996) have suggested that mineralisation at Cracow occurred during the Early Permian, associated with the emplacement of a large felsic intrusion underneath Cracow which was the source of the rhyolite dykes. U-Pb zircon dating of a rhyolite at Cracow gave an age of 291 ± 5 Ma (unpublished data, C Perkins, cited by Dong and Zhou, 1996).

Six stages of quartz vein formation have been recognised in the Golden Plateau system (Cracow Mining Venture Staff et al., 1990). The first three involve silicification of wallrocks and fracture filling but are essentially barren. Phase four involved brecciation of earlier quartz and wallrock, and rehealing by quartz precipitation commonly with colloform and banded textures. Most Au ore is associated with the fifth phase, during which all earlier precipitates were brecciated and rehealed. Gold is present as discrete grains of native gold or electrum in a sulphide or silicate matrix. Other minerals include sphalerite, chalcopyrite, pyrite, bornite and hessite, plus traces of arsenopyrite, marcasite, altaite, covellite, digenite and rutile (Cracow Mining Venture Staff et al., 1990). Propylitic, sericitic and intermediate argillic alteration assemblages are identified. The propylitic assemblage is located distal from the main lodes, whereas the sericitic and argillic assemblages are found within the main lode system and exhibit a vertical zonation from kaolinite-smectite±pyrite±hematite (intermediate argillic) in the upper levels to quartz-illite-smectite-sericite in the lower levels. The last phase of fine-grained quartz resealed quartz-breccias, is Au-mineralised, and probably reflects mixing of cool, acid, descending fluids with hot, near-neutral, upwelling hydrothermal fluids (Cracow Mining Venture Staff et al., 1990). Negative $\delta^{18}\text{O}$ values in vein quartz and altered wall rock suggests that the fluids responsible for the vein system were dominantly meteoric (Golding et al., 1987).

The Mount Terrible Volcanic Complex in north-central New South Wales lies within a belt of poorly exposed Permo-Carboniferous volcanic and intrusive rocks (Teale, 1998; Teale et al., 1999; Fig. 69). The volcanics and associated epithermal veins formed in a very Early Permian extensional setting overprinting a Late Carboniferous forearc (Stroud et al., 1999). Gold mineralisation was discovered in 1990 by Werrie Gold Ltd. and occurs in two areas of the Mount Terrible Volcanic Complex (Hillside and Silicon Valley deposits; Werrie Gold, 1996; Teale et al., 1998). Silicon Valley exhibits many characteristics ascribed to porphyry/breccia mineralisation styles, whereas Hillside has all the characteristics of gold-base metal sulphide-carbonate epithermal vein deposits (Werrie Gold, 1996).

The Hillside prospect (Werrie Gold, 1996; Teale, 1998; Teale et al., 1999) consists a number of subparallel, steeply NE-dipping lodes which trend 120° . A variety of alteration types have been recognised and Au is associated with late, low temperature epithermal carbonate-base metal sulphide veins and Bi-bearing sulphosalts (Werrie Gold, 1996). Higher Au grades are associated with reactivated and brecciated vein material (Teale, 1998; Teale et al., 1999). Galena, Fe-poor sphalerite, chalcopyrite, arsenian marcasite, pyrite and matildite are common in the upper domains of mineralised structures with chalcopyrite, arsenian pyrite, aikinite and berryite more common in the deeper zones. Dominant gangue phases are carbonate, quartz, chlorite, smectite, sericite, adularia, albite and zeolite (Werrie Gold, 1996; Teale et al., 1999). Mineralisation and alteration occurred over the temperature range of $150\text{--}350^\circ\text{C}$ based on mineral assemblages, fluid inclusion studies and sulphur isotope temperature estimates (Werrie Gold 1996; Teale, 1998).

The Silicon Valley Prospect is characterised by sheeted auriferous carbonate-base metal sulphide veins. These low temperature veins contain calcite, pyrite, low Fe sphalerite, chlorite, galena, and some quartz, with negligible to absent chalcopyrite. High temperature quartz-magnetite±pyrite±molybdenite stockwork

veins are adjacent to primary chalcopyrite-bornite±chalcocite intergrowths. Calcite-pyrite-tourmaline±rutile±quartz±chalcopyrite veins cut altered and brecciated volcanics and are considered to represent early high temperature (~350°C) veins (Werrie Gold, 1996). Moderate to intense hydrothermal alteration occurs through the Mount Terrible Volcanic Complex with early propylitic alteration widespread and potassic, phyllic and argillic alteration and variations of these alteration types noted (Teale et al., 1999).

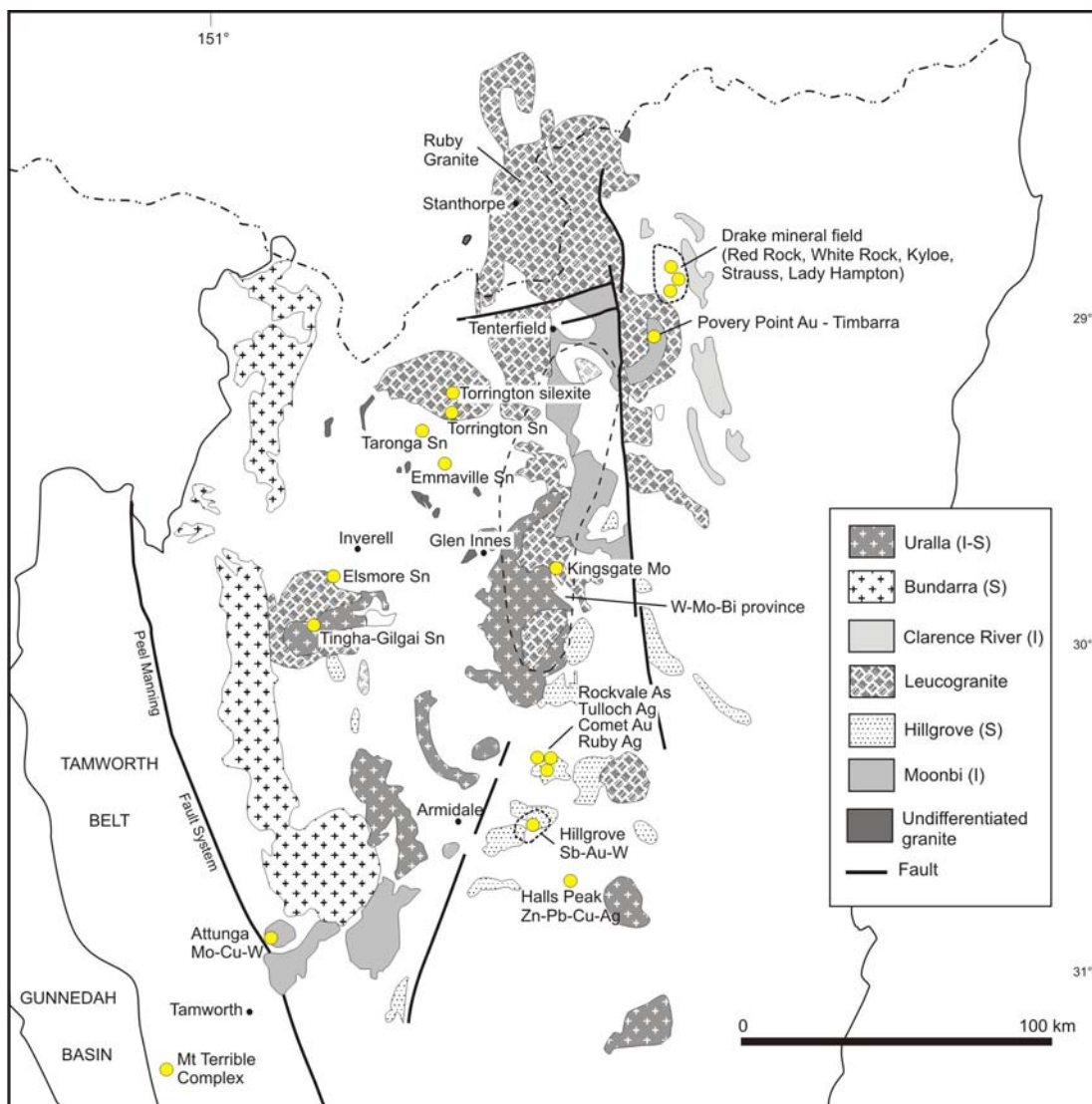


Figure 69. Simplified geological map of the southern New England Orogen showing granite suites after Shaw and Flood (1981) and the distribution of mineral deposits discussed in this report. Modified after Gilligan and Barnes (1990) and Solomon and Groves (2000).

3.5.4. Early Permian (~280 Ma) VHMS and related deposits, Mount Chalmers and Halls Peak

Early Permian rocks in the New England Orogen contain a number of small VHMS deposits, including Mount Chalmers in central Queensland (Fig. 68), and Halls Peak and Huntingdon in northern New South Wales (Fig. 69). Of these, the Mount Chalmers deposit is the most significant and has been described in greatest detail.

The Mount Chalmers deposit is hosted by the Chalmers Formation, which is constrained by U-Pb zircon ages from volcanoclastic rocks and rhyolites to ~277 Ma (Early Permian: Crouch, 1999). This unit, which forms part of the Berserker Group, is dominated by siltstone, volcanically derived sandstone and volcanic breccia as well as minor fossiliferous calcareous sandstone. This unit also contains rhyolite, some of which may be extrusive (Crouch, 1999). Crouch (1999) interpreted the maximum water depth of emplacement to be 200 m based on the fossil assemblage and the calcareous sandstone.

The Mount Chalmers VHMS deposit produced 1.14 Mt grading 1.99% Cu, 20.8 g/t Ag and 3.38 Mt Au with minor Zn and Pb (Taube, 1990). This deposit consists of two main massive sulphide lenses overlying a laterally extensive siliceous stringer zone that extends 50 m below the massive sulphide lenses. The deposit is virtually undeformed and contains well bedded and graded clastic sulphides near the top of the ore lenses (Large and Both, 1980). The main ore minerals are pyrite and chalcopyrite with variable sphalerite and galena, and trace Ag sulphosalts (Large and Both, 1980). This deposit is unusual in a number of characteristics, including the high Au grades, the inferred shallow water environment of deposition (Sainty, 1992), and the presence of kaolinite as a significant gangue (McLeod, 1985). The latter characteristic suggests that the Mount Chalmers deposit fits into the high-sulphidation sub-class of VHMS deposits. In addition the Mount Chalmers deposit, Taube and van der Helder (1983) described a number of small VHMS prospects in the Chalmers Formation, and Crouch (1999) described a number of small lode gold deposits of unknown age to the south of the Mount Chalmers deposit.

Apparently correlative rocks of the Early Permian age in the southern New England Orogen in New South Wales also hosted small VHMS prospects at Halls Peak (Gilligan and Barnes, 1990). The Halls Peak Volcanics host a number of small VHMS deposits which historically have produced very small tonnages (16,000 t in total) of ore, though at very high grades (31-32% Zn, 19-21% Pb, 1.0-2.5% Cu and 900-1166 g/t Ag; Moody et al., 1993). The ores comprise 50-100% sulphide minerals and include sphalerite, galena with subordinate pyrite and chalcopyrite and trace tetrahedrite and arsenopyrite. The massive sulphide lenses are underlain by a zone of quartz-sericite-pyrite alteration assemblages associated with quartz-sulphide-carbonate stringer veins (Moody et al., 1993). Other deposits of similar age and origin include Silver Spur (past production totalled 68.3 t Ag and 0.14 t Au from 0.09-0.1 Mt of ore with JORC-compliant resources of 0.808 Mt grading 3.56% Zn, 1.25% Pb, 0.17% Cu, 2.25 opt Ag and 0.09 g/t Au [Murray, 1990, <http://www.macmin.com.au> (accessed 23 August 2008)]) in southern Queensland and Huntingdon in New South Wales (Gilligan and Barnes, 1990).

Rocks from both the Chalmers Formation and the Halls Peak Volcanics have been interpreted to have formed upon thinned continental crust associated with subduction. Crouch (1999) interpreted the Berserker Subprovince, which contains the Mount Chalmers deposit, as either a back-arc or intra-arc extensional basin formed on a continental margin, whereas Moody et al. (1993) interpreted the Halls Peak Volcanics based on the geochemistry to have formed upon thinned continental crust in an overall subduction-related environment. Crouch (1999) inferred that the Mount Chalmers deposit formed along the eastern margin of a back-arc basin, proximal to the continental arc, with the Bowen Basin forming the bulk of the back-arc basin. This setting is consistent with that inferred for the Cambrian Mount Lyell deposits in Tasmania, suggesting that the eastern margin of the Bowen Basin has potential Cu-Au-rich VHMS deposits.

3.5.6. Lode gold deposits, north Queensland

Minor Carboniferous to Permian lode gold deposits occur in north Queensland. The most significant are these are in the Hodgkinson goldfield, to the northwest of Cairns (Fig. 67), which produced about 9 t of Au and over 3 t of Sb (Murray, 1990). This goldfield comprises a number of separate mineralised zones – Beaconsfield, Union, Thornborough, Kingsborough and Northcote – hosted by metasedimentary rocks of the Hodgkinson Formation and localised close to major north-west trending structures (Vos and Bierlein, 2006). Gold mineralisation is vein hosted with sulphide-poor (arsenopyrite-pyrite±sphalerite-galena-tetrahedrite-stibnite) Au-Sb quartz veins, with earlier gold-quartz veins and later gold-stibnite-quartz veins (Peters et al., 1990; Vos and Bierlein, 2006). Fluid inclusion data (Peters et al., 1990; Vos and Bierlein, 2006) indicate low salinity fluids with minor CO₂, and temperatures of 150-400°C, consistent with

lode/orogenic gold mineralisation styles, though Peters et al. (1990) did not rule out a distal magmatic contribution. The timing of mineralisation is uncertain. Morrison and Beams (1995) suggested a Carboniferous age of ~ 328 Ma, whereas Davis et al. (2002) suggested much of the lode gold in the Hodgkinson was Permian in age, and Vos and Bierlein (2006) proposed two episodes of mineralisation related to successive deformation events in the Late Devonian-early Carboniferous and the middle Carboniferous. In proposing young (Permian) gold, Davis et al. (2002) have suggested that whilst quartz veins are of multiple ages, the gold mineralisation is one age, related to the last major orogenic deformation event in the province. What is not in dispute is the structural control and relationship of the gold mineralisation to major deformation, so multiple gold events within a region or within the one deposit are perhaps not surprising. The best example of this is the Amanda Bel goldfield in the Broken River Province which in addition to gold mineralisation in the Early and Late Devonian also apparently contains a mid to Late Carboniferous mineralising ($\text{Sb}\pm\text{Au}-\text{As}$) event (e.g., Teale et al., 1989; Vos et al., 2005).

The Palmer River gold field (northwest of Cairns) produced over 40 t Au, but this was dominantly alluvial, with only minor lode gold (Murray, 1990). Like the Hodgkinson goldfield, the sources of the Palmer River alluvials may be of Permo-Carboniferous age (Morrison and Beams, 1995; Davis et al., 2002), but they could also be related to Devonian deformation (Murray, 1990).

3.5.7. Lode gold-antimony deposits, southern New England Orogen

The Hillgrove mineral field (Fig. 69) is one of the largest Au producers in NSW and is Australia's only export Sb producer. The structurally-controlled Au-Sb quartz vein(-breccia) systems were described by Boyle and Hill (1988) and Boyle (1990) and over 200 lodes have been exploited since 1877 (Boyle, 1990). These lodes are localised in a northwest-striking belt about 5km and 2km wide (Ashley et al., 1994; Ashley and Creagh, 1999), with production of 25 t Au and over 60 kt of stibnite. The veins and breccias are hosted in metamorphosed (biotite-grade) Carboniferous flysch-type sedimentary rocks (fine-grained greywacke, siltstone, siliceous argillite of the Gurrakool beds), late Carboniferous-Early Permian (~ 300 Ma) S-type granites of the Hillgrove Suite and the Early Permian Bakers Creek I-type diorite-tonalite-granodiorite complex (Boyle, 1990; Collins et al., 1993; Landenberger et al., 1993).

The Hillgrove mineral field shows a strongly developed pattern of N and NW trending shears. The majority of the veins occur in or near these shear zones, and in zones of intense brecciation. Lode structures transgress intrusive contacts between granites and metasedimentary rocks (Boyle, 1990). Three distinct veining episodes occur within the field: a barren early quartz episode, the mineralising episode and a late barren quartz-calcite-chlorite episode. The mineralised veins have four phases: quartz-scheelite; quartz-arsenopyrite-pyrite-gold, quartz-stibnite-gold-silver and quartz-stibnite-calcite. The principal vein minerals are quartz, pyrite, arsenopyrite, stibnite, scheelite, calcite, and gold. Minor amounts of jamesonite, stannite, sphalerite, galena, chalcopyrite, pyrrhotite, marcasite, tetrahedrite, antimony, cassiterite and wolframite, among other phases, have also been identified (Boyle, 1990).

Calc-alkaline (shoshonitic) lamprophyre dykes have a close spatial association with the auriferous lodes. Dilational lode structures acted as conduits for dyke intrusion, which occurred before and after major quartz-stibnite veining. Field relations constrain the age of emplacement of lamprophyre dykes and Au-Sb lodes as post-dating deformation of the Hillgrove Suite (Ashley et al., 1994). Ashley et al. (1994) provided phlogopite K-Ar ages for lamprophyres of 255 ± 5 Ma and 257 ± 5 Ma, ages consistent with K-Ar ages of lamprophyres from Rockvale, 20km north of Hillgrove (Kent, 1994). While the mineralising and lamprophyre dyke events are coeval with intrusion of high-K I-type granites (Boyle, 1990; Ashley et al., 1994; Kent, 1994), there is no direct genetic connection between the intrusions and the Au-Sb mineralisation. The mineralisation displays shallow mesozonal characteristics, and may be related to structural focussing of hydrothermal fluids derived largely from metamorphic devolatilisation reactions (Ashley et al., 1994). The close temporal and spatial relation of dykes, mesothermal Au-Sb veins and I-type intrusions are interpreted to be manifestations of the post-collisional setting and influx of mantle-derived heat and partial melts into the New England Orogen during the Permian (Ashley et al., 1994).

Hydrothermal Ag-As-base metal, Au-(W-Bi) and Sb-(Au) vein mineralisation in the Rockvale region (Fig. 69) around the Rockvale Adamellite either post-dated or was synchronous with dyke emplacement at ~255 Ma (Kent, 1994). While Gilligan and Barnes (1990) suggested veins of the Comet Au, Tulloch Ag, Ruby Ag and Rockvale As deposits around the Rockvale Adamellite may have formed during hydrothermal activity associated with this granite, it appears more likely that Ag-As-base metal mineralisation is the product of fluids evolved by a felsic granitoid similar to those spatially related to the mineralisation (Kent, 1994); the fluids, however, were considered by Comsti and Taylor (1984) to be of metamorphic hydrothermal origin.

3.5.8. Drake Mineral Field gold and base metal deposits

The Drake Volcanics (Fig. 69) comprise a complex, andesite-dominated, calc-alkaline suite of lavas, tuffs and epiclastic rocks together with subvolcanic, comagmatic stocks and dykes of Late Permian age (Bottomer, 1986) that host numerous epithermal Au, Ag, and base metal deposits over an area of about 300 km². These deposits occur as veins, stockworks, stratabound and lenticular disseminations, and breccia infillings (Bottomer, 1986; Perkins, 1987, 1988; Brown et al., 2001), locally associated with porphyritic andesitic, dacitic and rhyolitic intrusives and lavas. All the major deposits lie within the Drake Volcanics, favouring a close genetic relationship between volcanism and mineralisation (e.g., Mount Carrington epithermal Au-Ag-Zn deposits: Brown et al., 2001), although the actual age of mineralisation is unknown. The deposits are regarded as being of (sub)volcanic epithermal origin, and although not of great commercial significance, taken together, they make up an unusual metal province (see Perkins, 1988).

Mineralisation and pervasive hydrothermal alteration were contemporaneous with sedimentation and volcanism (Bottomer et al., 1984; Bottomer, 1986; Perkins, 1988; Brown et al., 2001). The primary mineralisation consists of Ag sulphosalts and native metal alloys, with associated base metal sulphides and pyrite, generally contained within silicate±pyrite alteration zones. Both mineralisation and alteration are multistage, typically with an early barren or low-grade sericite-clay assemblage overprinted by a later ore-associated quartz-K-feldspar assemblage. Intense sericite-quartz-carbonate assemblages and silicified zones associated with mineralised rock grade outwards into propylitic assemblages. Permeable host rocks and the presence of a structural feeder zone are features common to all the deposits (Bottomer, 1986).

The Red Rock deposit is one of a number of epithermal Ag-Au occurrences in the Drake Volcanics (Perkins, 1987, 1988), and is hosted in hydrothermally brecciated volcanoclastic rocks of the Cataract River Member. Mineralisation consists of disseminations and vein stockworks of precious and base metal lodes, accompanied by quartz, calcite, adularia, pyrite, illite, illite-montmorillonite, chlorite and sphene alteration. According to Perkins (1987, 1988), seawater may have been the dominant component of the ore fluids, while the close temporal and spatial relationship to comagmatic intrusives suggests that they may have been responsible for heating the terrane and/or contributing directly to the ore fluids (Perkins, 1988). A submarine environment of formation for the Red Rock deposit implies that the Drake mineralisation may have similarities to volcanogenic massive sulphide or Kuroko-type deposits, the latter of which form in marine settings in similar tectonic overall environments to epithermal deposits (Perkins, 1988).

3.5.9. Early to Middle Triassic (250-235 Ma) granite-related deposits in the southern New England Orogen

Although, with the exception of tin, relatively minor in terms of metals produced, the Early to Middle Triassic mineralising epoch in the New England Orogen contains a very diverse metallogenic assemblage. Blevin and Chappell (1996) characterised the northern New England Orogen as a Cu-Mo-Au province and the southern New England Orogen (mainly in NSW) as a Sn-Mo-W-polymetallic province. Very latest Permian to Triassic magmatism in the New England Orogen is directly and indirectly responsible for the majority of mineral deposits.

Triassic magmatism in the New England Orogen was responsible for the generation of major Sn fields in the southern New England Orogen at Stanthorpe, Emmaville-Torrington and Tingha-Elsmore, with production of approximately 65, 90 and 70 kt of cassiterite concentrates derived from these fields, respectively (Kleeman, 1990). The historically important Sn deposits (cassiterite-bearing disseminated greisen, greisen-bordered vein, sheeted vein and pegmatite) are associated with the Ruby Creek, Mole, Gilgai, and Elsmore leucogranites in the southern New England Orogen (Ashley et al., 1996). The Mole Granite is the most intensely mineralised granite in the southern New England Orogen (Henley et al., 1999).

The Late Carboniferous to Triassic New England Batholith, with an outcrop area of $\sim 15000\text{km}^2$, is comprised of both synorogenic, Late Carboniferous to Early Permian, peraluminous S-type granites and post-orogenic, Permo-Triassic I-type intrusions. Shaw and Flood (1981) used mineralogical, geochemical, isotopic and age criteria to subdivide the New England Batholith into the Bundarra (S-type Late Carboniferous-Early Permian), Hillgrove (S-type Late Carboniferous-Early Permian), Moonbi (magnetite-bearing I-type Permo-Triassic), Uralla (I-type, Permo-Triassic) and Clarence River (magnetite-bearing I-type, Permo-Triassic) suites (see Fig. 69). Chappell (1994) later renamed these supersuites. Several economically significant and highly fractionated felsic granites were assigned to a separate 'leucoadamellite' group, subsequently reclassified as leucomonzogranite and incorporated into the Moonbi Supersuite (Triassic; Chappell, 1994).

The felsic granites of the Stanthorpe Supersuite (including the Mole Granite) possess high concentrations of K, Rb, Sr, Ba, U, Th and Pb and are relatively depleted in Ba, Sr, Eu and LREE (Blevin and Chappell, 1993). The Stanthorpe Supersuite is by far the most important group of mineralised granites in the region. Deposits occur in various forms including veins, stockworks, pipes, disseminations, greisens and skarns (Gilligan and Barnes, 1990). Over 2000 low tonnage occurrences are concentrated within and around the margins of some of these granites and include Sn, W, Mo, Ag, As, Bi, Cu, Pb, Au, fluorite, beryl and topaz mineralisation. The mineralisation related to leucocratic granites can be divided into Mo, Sn, polymetallic vein and Au associations (Blevin et al., 1996; Stroud et al., 1999; Blevin and Chappell, 1992, 1995; Blevin, 2004).

3.5.9.1. Deposits related to the Ruby Creek Granite

Triassic (245-238 Ma) high-K granites of the Tenterfield-Stanthorpe region form a distinct group within the I-type Moonbi Supersuite (Stanthorpe Supersuite of Blevin and Chappell, 1996; Fig. 69) and have recently been split off into the Stanthorpe Supersuite (Donchak et al., 2007). Three types of granites are recognised within the Stanthorpe Supersuite: (1) the 248-243 Ma Bungulla type, (2) the 242-238 Ma Stanthorpe type, and (3) the 241-239 Ma Ruby Creek type (see below). The Stanthorpe Supersuite crops out in two parts – the northern Stanthorpe mass and the southern Timbarra mass. The Timbarra mass hosts the Timbarra Au deposits (see below). Within the Stanthorpe mass, the Ruby Creek Granite (Fig. 69) hosts Mo, W and Sn mineralisation (Blevin and Chappell, 1996), as outlined below.

Southwest of Stanthorpe (Fig. 69), the Carboniferous Texas beds have been intruded by the Ruby Creek Granite (Re-Os molybdenite age of 242.1 ± 0.7 Ma; Norman et al., 2004), which extensively intrudes, the Stanthorpe Granite (Donchak et al., 2007). Parts of the Ruby Creek Granite are very strongly fractionated as in the Sugarloaf, Kilminster and Sundown areas where it is associated with Sn, Mo, W, Bi, base metal, and Au mineralisation (Denaro and Burrows, 1992; Blevin and Chappell, 1992). Denaro and Burrows (1992) described the vast array deposit types genetically associated with the Ruby Creek Granite into the following:

- a) Siliceous highly fractionated phases of the Ruby Creek Granite containing wolframite-molybdenite-pyrite-minor bismuth as disseminated and vein mineralisation.
- b) Mo- and B-bearing quartz pipes in the marginal Ruby Creek Granite phase, closely related to deposits in (a).

- c) Flat-lying and near-vertical quartz greisen veins, pegmatite veins and greisen and pegmatite pods in the marginal porphyritic phase of the Ruby Creek Granite. Ore minerals include cassiterite, wolframite and molybdenite with minor arsenopyrite, chalcopyrite, sphalerite and galena. In many places, the quartz greisen veins occur as sheeted vein and stockwork deposits which extend into the surrounding country rock.
- d) Greisen pipes occurring as zones of intense alteration along the intersection of quartz greisen veins in stockwork deposits in the Ruby Creek Granite.
- e) Quartz-arsenopyrite lodes along faults and sheared zones in the Ruby Creek Granite and adjacent Texas beds at Jibbinbar.
- f) Joint-controlled sheeted vein systems extending from the Ruby Creek Granite to ~200 m above the apices and flanks of granite cupolas. Veins in the Texas beds consist of quartz-arsenopyrite-cassiterite-wolframite-topaz-muscovite. Veins in the Stanthorpe Granite consist of quartz-muscovite-cassiterite-wolframite-molybdenite-Bi and show pervasive greisenisation.
- g) Quartz-arsenopyrite-cassiterite-chalcopyrite lodes as rich, lenticular deposits within sheeted veins systems in the Texas beds. The lodes extend from the contact zone up to 150m above the apices and flanks of cupolas of Ruby Creek Granite. There is a general vertical zonation from a Sn-As zone near the granite contact, through a Cu-Sn-As zone, to a higher level Sn-As zone. At least two generations of hydrothermal fluid release and ore deposition are inferred.
- h) Base metal sulphide mineralisation in porphyry and other dykes, interpreted as being genetically related to the Ruby Creek Granite.

The south-western extension of the Severn River Fault Zone contains the Sn and polymetallic deposits of the Sundown area (Fig. 68), which have been historically worked for Cu, Ag and As and minor Au, Mo and W (Donchak et al., 2007). Joint controlled and sheeted quartz-cassiterite and quartz-arsenopyrite-cassiterite vein systems of the Sundown Prospect are hosted within hornfelsed Texas beds capping very shallow-level plutons of Ruby Creek Granite, which supplied the mineralising fluids. The veins are structurally controlled and mineralisation is independent of lithology. The veins are 0.5 to 10 mm wide and are irregular in distribution. Economic Sn grades are restricted to 10 to 12 m wide zones of intense fracturing (Denaro and Burrows, 1992). Alteration haloes adjacent to veins consist of 10 to 20mm of silicified and sericitised wallrock. At the Sundown Prospect, mineralised zones were in the order of 100 to 200 m long and extended down to the Ruby Creek Granite at approximately 80m depth (Denaro and Burrows, 1992).

The partially unroofed upper surface of the Ruby Creek Granite at Sundown coincides with zones of intense veining in the hornfelsed sedimentary rocks (e.g., Goode et al., 1982). Mineralisation in the granite has produced a greisen assemblage of quartz-muscovite-topaz-fluorite±cassiterite±chlorite±siderite. The ore veins contain two assemblages – an earlier cassiterite-muscovite-quartz-topaz-fluorite-beryl±arsenopyrite assemblage and a later chalcopyrite-pyrrhotite-sphalerite-chlorite-carbonate assemblage. A shallow depth for vein formation (1-3km) is indicated, as pressures were sufficiently low to prevent boiling in the CO₂-poor Sundown system, while groundwater involvement was likely after the main Sn mineralisation (Solomon and Groves, 2000).

North-west of the Sundown Prospect, the Jibbinbar Granite is associated with significant As mineralisation and contains small shear-hosted As, Pb, Ag, Zn and Cu deposits (Donchak et al., 2007). Arsenopyrite is concentrated in quartz veins and quartz-rich lodes distributed along east to north-east-trending shear zones and/or faults in the granite (Denaro and Burrows, 1992). Silver and Au also occur in arsenopyrite lodes at Jibbinbar, but grades are erratic (up to 5g/t; Denaro and Burrows, 1992; Donchak et al., 2007). In contrast to the Ruby Creek Granite at Sundown, the Jibbinbar Granite is not associated with significant Sn mineralisation, with available data indicating the Jibbinbar Granite is not as highly fractionated as, and is chemically and mineralogically distinct from, the Ruby Creek Granite (Donchak et al., 2007).

3.5.9.2. Timbarra intrusion-related gold deposits

The Timbarra Au deposits are located about 15 km ESE of Tenterfield and 25 km SW of Drake in northeastern NSW (Fig. 69). Alluvial and colluvial Au were discovered at Timbarra in 1850 and primary

mineralisation at Poverty Point in the 1870s (Cohen and Dunlop, 2004). Timbarra is located within a Palaeozoic subduction-related accretionary complex of oceanic crustal terranes and is dominated by I-type, high-K Permo-Triassic granites of the New England Batholith (Gilligan and Barnes, 1990; Mustard et al., 1998). The Timbarra mass of the I-type Stanthorpe Suite hosts the Timbarra Au deposits (Cohen and Dunlop, 2004).

The Timbarra Au deposits represent an economically significant and distinctive member of the intrusion-related class of Au deposits (Mustard, 2001, 2004; Thompson et al., 1999; Lang et al., 2000). The five known deposits possess a total identified mineral resource of 16.8 Mt at 0.73 g/t Au, for a total of 12.3 t (Ross Mining NL 1998 Annual Report). The Au deposits are found within the upper levels of highly fractionated plutons, stocks and dykes of the Stanthorpe Granite. Disseminated ore, present in all five deposits, comprises >95% of the overall resource and occurs predominantly as gently-dipping tabular to lenticular bodies that are localised beneath fine-grained aplite carapaces and internal aplite layers. Disseminated ore consists of Au-bearing muscovite-chlorite-carbonate altered granite and infill of primary miarolitic cavities within massive leucomonzogranite (Mustard, 2001). Structurally-controlled ore forms the remaining 5% of the Timbarra resource, and comprises minor vein-dykes and quartz-molybdenite veins. Pegmatite sheets in the upper portion of the granites have been mined extensively for Au which occurs predominantly along microfracture arrays within pegmatite quartz (Simmons et al., 1996). Pervasive sericite-chlorite-albite alteration assemblages are associated with the ores, with minor quartz or carbonate veining. The Au-mineralised zones are localised and enhanced by faults, joints and cooling fractures within the roof of the granite (Mustard et al., 1998).

Mineralisation and alteration share a common paragenetic sequence of precipitation, as outlined by Simmons et al. (1996) and Mustard (2001). Quartz, K-feldspar, minor biotite and albite are the earliest and most abundant phases, commonly lining primary cavities and vein-dykes. Subsequent minerals include coeval arsenopyrite, pyrite, fluorite and molybdenite. The last phases to form include muscovite, chlorite, gold, calcite, Ag-Bi telluride, Pb-Bi telluride, plus rare galena and chalcopyrite. The Au ore has a low total sulphide mineral concentration (<1%) (Simmons et al., 1996; Mustard, 2001). The two elements most closely associated with Au are Mo and Bi (Cohen and Dunlop, 2004), with association also with Sb and As (Simmons et al., 1996). Ore contains elevated concentrations of Bi, Ag, Te, As, Mo and Sb (Mustard, 2001).

The Poverty Point deposit (containing 1.90 t of Au, roughly 15% of the current resource; Mustard, 2001) is constrained beneath the aplite carapace of the Stanthorpe Granite. The orebody is flat-lying to shallowly west-dipping and tabular in form. Its orientation reflects the geometry of the bounding aplite carapace, and it is less than 50 m thick, 400 m long and 200m wide (Cohen and Dunlop, 2004). Rare veins within the granite at Poverty Point contain quartz, molybdenite, gold, pyrite, calcite, albite, fluorite and illite, and are interpreted to represent limited fracturing during development of the mineralised zones within the granite. Disseminated gold in the granite contains 5-15 wt% Ag, and has a close spatial association with bismuthinite and molybdenite (Simmons et al., 1996). Primary fluid inclusions in quartz texturally associated with Au consist of CO₂-rich and low salinity H₂O-rich types (Simmons et al., 1996). Blevin (2002) reported a Re-Os age of 248.7 ± 0.5 Ma (weighted average of 2 samples) for coarse molybdenite from a quartz-molybdenite-chalcopyrite vein in Stanthorpe Granite at Poverty Point.

Mineralisation at Timbarra is related to highly fractionated high-level granites with Au deposition being one of the last events (Mustard et al., 1998). Mineralising and altering fluids were pooled beneath microgranite sills or layers and carapaces to form broad lensoid and tabular zones (Mustard et al., 1998). This style of mineralisation, principally disseminated with minor sheeted veins, is analogous to that displayed by granite hosted W and Sn deposits (Simmons et al., 1996). The oxidised host granite, low sulphide ores, Au-Bi-Ag-Te-Mo geochemical signature, muscovite-chlorite-carbonate alteration assemblage and low-salinity aqueous and carbonic fluids suggest that Timbarra is part of the intrusion-related gold deposit class (cf. Thompson et al., 1999), although is unusual in having a disseminated style of mineralisation (Mustard, 2001, 2004).

3.5.9.3. Deposits associated with the Mole and Gilgai Granites

Tin-tungsten deposits extend from the northern part of the New England area down to about latitude 31° in a broad arc which also includes a north-south belt of W±Mo±Bi deposits associated with leucogranites stretching from Kingsgate to Tenterfield (Solomon and Groves, 2000; Henley et al., 1999; Fig. 69). Historically important Sn deposits are associated with southern New England Orogen leucogranites in the Stanthorpe, Emmaville-Torrington and Tingha-Elsmore regions (Ashley et al., 1996; Fig. 69).

Most of the over 200 kt of Sn concentrate has come from alluvial and placer deposits shed from the granites, the most important of these being the Mole Granite. The Mole Granite is only partially unroofed, with ~650 km² of exposed surface and another ~1200 km² still buried under sedimentary and volcanic country rocks (Kleeman, 1982). Its age has been determined by K-Ar, Ar-Ar, and Rb-Sr dating of whole rocks and hydrothermal vein minerals at 246 ± 2 Ma (Kleeman et al., 1997), confirmed by recent U-Pb magmatic zircon and monazite ages of 247.6 ± 0.4 and 247.7 ± 0.5 Ma respectively, with hydrothermal xenotime giving a slighter younger age of 246.2 ± 0.5 Ma (Schaltegger et al., 2005; Pettke et al., 2005). The Mole Granite is part of the group of Permo-Triassic I-type plutons, fractionated I-type leucocratic granites, which also includes the Gilgai, Stanthorpe and Ruby Creek Granites (Henley et al., 1999; Fig. 69).

Over 1200 identified mineral deposits and occurrences are genetically related to the Mole Granite (Henley et al., 1999), most of which having been mined for either Sn or W/Bi, and a smaller number for Cu, Pb, As, Zn, Ag, or Au. It has produced over 89 kt of cassiterite and has been a significant producer of W, As, Ag, and emeralds. It also hosts the world's largest silexite (quartz-topaz greisen) deposit (Henley et al., 1999). There is a prominent metal zonation from Sn-dominated deposits (±B, Cu, Pb, Zn) within the granite, through W-dominated deposits (±F, Be, Bi, As, Mo) at the granite margin, to sulfide-rich polymetallic deposits (As, Pb, Zn, Cu, Sb, Ag) in the surrounding sediments (see Henley et al., 1999).

Except for some large, mineralised fault systems, all ore deposits are confined to within 100 to 200 m vertical distance from the granite contact (Kleeman et al., 1997), and more deeply eroded parts of the granite are not mineralised. Mineralisation within the granite occurs mostly in quartz veins, or as disseminations in massive quartz-topaz greisens (silexites) at the granite margin. Vein deposits within the Mole Granite are mainly steeply dipping (apparently formed in joints) with a common assemblage of quartz-chlorite-sericite-cassiterite with rare fluorite. Cassiterite tends to be associated with chlorite. The other main vein assemblage is quartz-cassiterite. A few wolframite-bismuth veins are present and some Sn veins carry wolframite. A few base metal sulphide-rich assemblages are also present. Late quartz-adularia veins cut cassiterite and wolframite ores throughout the granite (Kleeman et al., 1997; Henley et al., 1999).

Ore deposits outside the granite include pegmatites, sheeted veins (above ridges or cupolas in the roof), and mineralised faults and shear zones (Aude'tat et al., 2000a, b). Small veins and sheeted vein systems outside the granite were grouped by Weber (1975) into cassiterite, cassiterite-arsenopyrite and metal sulphide (galena, sphalerite, chalcopyrite and pyrite) vein types. Most veins contain quartz, cassiterite and arsenopyrite. Large multiple (sheeted) vein systems constitute bulk (economic) deposits, as at Taronga, northwest of Emmaville, where a resource of over 46 Mt of more than 0.14% Sn (Kleeman et al., 1997) has been defined in hornfelsed sediments. The deposit consists of a sparse stockwork of narrow cassiterite-bearing veins overlying a buried granite cupola (Kleeman et al., 1997; Gilligan and Barnes, 1990).

Tungsten-dominant mineralisation accounts for ~25% of known occurrences associated with Mole Granite (Henley et al., 1999). Most of these occurrences are in the Torrington pendant (Fig. 69) where W-Bi mineralisation is found disseminated within silexite or within multiple narrow veins within the silexite. Tungsten occurs as wolframite; W mineralisation within the granite occurs as discrete quartz veins or multiple veins (cm up to 1.5m wide) associated with cassiterite or monazite (Henley et al., 1999).

In the Tingha-Inverell area, the Late Permian Gilgai Granite intruded the Early Triassic Tingha Adamellite (Uralla Supersuite) and Permian to Carboniferous country rocks. Hydrothermal fluids related to the Gilgai Granite have emplaced base metal and joint controlled quartz-chlorite-sericite-cassiterite greisen vein

deposits along pre-existing joints in the upper reaches of the Gilgai Granite and Tingha Adamellite. Over 70 kt of Sn has come from the Tingha-Gilgai area (Gilligan and Barnes, 1990). Of the ~80 base metal deposits recognised in this granite, most are vein and dissemination type deposits associated with chloritic alteration. Mineralisation includes Pb-Zn-Ag-Cu-As sulphides, with minor Mo and Sn. Within the >200 cassiterite deposits, mineralisation is dominantly cassiterite, with minor to rare As, Fe, Cu, W, and Mo sulphides and/or oxides (Brown and Stroud, 1993).

3.5.9.4. Kingsgate W-Mo pipes

Molybdenum deposits are commonly associated with relatively oxidised, mesocratic to leucocratic granites of the I-type Moonbi Supersuite and important Mo-Bi-W deposits occur as pipes, veins and disseminations at the outer contact zones of the Kingsgate Granite (Weber et al., 1978; Ashley et al., 1996).

The Kingsgate Mo±Bi pipes east of Glen Innes in New England (Fig. 69) have yielded approximately 350 t of molybdenite and 200 t of Bi from over 50 pipes (Weber, 1975; England, 1985). About 70% of these pipes are known from a 3 km long, north-northwest-trending belt. These pipes occur in clusters within and near the top of small, shallow plutons of greisenised leucocratic, strongly fractionated, magnetite-bearing I-type granites, which intrude a hornblende-biotite granite and sedimentary rocks (Weber, 1975). The pipes vary from 15m to more than 150 m in length and in diameter from 1 to 20m, but most commonly they are 2.5 to 8 m in diameter, extending down dip for more than 80 m. The pipes contain a wide variety of minerals including molybdenite, bismuth, bismuthinite, gold, wolframite, pyrrhotite, pyrite, chalcocopyrite, arsenopyrite, galena and cassiterite (Gilligan and Barnes, 1990), plus silver, joseite, cosalite and bismutite (Weber et al., 1978). They have a core of vein quartz, flanked by a high silica zone that grades out to altered granite. Foliated masses of molybdenite and separate masses of bismuth and bismuthinite occur in the quartz core. Pyrrhotite with chalcocopyrite, bismuth, sphalerite and galena is prominent locally, with sporadic cassiterite and wolframite (Weber et al., 1978). $\delta^{34}\text{S}$ values of molybdenite are close to zero per mil (Herbert and Smith 1978). The origin of the pipes is unknown, although Weber et al. (1978) suggest they may have been formed by hydrothermal fluids expelled from the crystallising magma. Similar W-Mo-Bi pipe systems in eastern Australia are rarely restricted in occurrence to elsewhere in New England, at Bamford Hill and Wolfram Camp in north Queensland and at Whipstick in the Bega Batholith of southeast New South Wales.

3.5.9.5. Attunga skarn deposits

Moonbi Supersuite granites are also associated with scheelite skarns. The scheelite-molybdenite-bearing andradite skarn at Attunga (Fig. 69) lies adjacent to a high-K intermediate Triassic intrusion, with major exoskarn replacement of marble and calc-silicate hornfelses and minor endoskarn replacement of the quartz monzonite intrusive (Ashley et al., 1996). The Attunga skarn displays oxidised mineralised assemblages (andradite, magnetite, pyrite, chalcocopyrite, bornite) with associated minor Bi, W and Mo minerals (Ashley et al., 1996).

While several skarn deposits are located adjacent to Triassic granites in the southern New England Orogen, they are not abundant, mainly due to the paucity of suitably reactive hosts (e.g., carbonate). Nevertheless, Ashley et al. (1996) suggests the Attunga region hosts potential for further Cu-Au-W-Mo skarn deposits. According to Ashley et al. (1996), more distal replacements of carbonate and organic-bearing sequences to form the subtle 'Carlin-type' Au deposits are also possible targets in this region.

3.5.10. Late Permian to Middle Triassic (260-235 Ma) porphyry copper±molybdenum±gold and related deposits, central New England Orogen

As discussed by Blevin and Chappell (1996) the Late Permian to Middle Triassic granite-related deposits in the central New England Orogen are characterised by a dominant Cu±Mo±Au metallogenic assemblage,

which contrasts with the Sn-W-dominant assemblage in the southern New England Orogen. Numerous large-scale hydrothermal systems with fracture-controlled and disseminated $\text{Cu}\pm\text{Mo}\pm\text{Au}$ mineralisation of the type characteristic of porphyry ore deposits occur in the northern New England Orogen (Horton, 1982; Ashley et al., 1996). The Queensland part of the Permo-Triassic magmatic belt is characterised by low grade, high level, porphyry Cu-Mo deposits (e.g., the Moonmera, Coalstoun and Anduramba deposits; Fig. 68) and base metal and Cu-Au skarns (e.g., Biggenden, Glassford Creek, Many Peaks, Ban Ban; Fig. 68), many of which were reviewed by Horton (1982). Most are of Triassic age and are associated with diorite, tonalite and granodiorite porphyries with zoned potassic, phyllic and propylitic alteration. Emplacement of porphyry intrusions has been influenced by regional-scale lineaments and the north-northwest Yarrol Fault appears to connect a considerable number of porphyry Cu and porphyry Mo stockwork deposits and base metal and Cu-Au skarn deposits (Ashley et al., 1996).

The Coalstoun Lakes prospect is a significant subvolcanic porphyry Cu-Au-Mo system in strongly brecciated and altered Good Night beds and Early to Middle Triassic diorite-monzonite stocks (Denaro et al., 2007). The prospect has an inferred resource of 85.58Mt at 0.287% Cu to 300m depth for a contained 245 615t of Cu (Metallica Minerals Ltd, quoted in Denaro et al., 2007). The geology and mineralisation at this prospect have been described in detail by Ashley et al. (1978) and Solomon and Groves (2000).

At the Coalstoun Lake deposit, Late Permian tonalitic and dioritic plutons were intruded into Devonian-Carboniferous siltstone, cherts and greywackes of the Curtis Island Group (Ashley et al., 1978; Denaro et al., 2007). In the central part of the deposit, the pluton of porphyritic biotite microtonalite is ringed by breccia pipes that have gradational boundaries either to intrusive or sedimentary rocks. The pluton has a core of biotite alteration which grades upward to quartz-sericite-pyrite assemblage and outward to a chlorite-carbonate assemblage. The biotite zone has pervasive hydrothermal biotite and quartz with minor anhydrite, feldspar, magnetite, rutile, haematite and fluorite and the veins of the stockwork contain combinations of quartz, pyrite, chalcopyrite, magnetite, anhydrite, K-feldspar and biotite. The chlorite-carbonate zone contains pervasive chlorite, calcite, and minor epidote, albite, sericite, clay, tourmaline, rutile, sphene, and haematite. The vein minerals are chlorite, carbonate, pyrite, gypsum and haematite (Ashley et al., 1978; Denaro et al., 2007). The highest Cu grades are centred on the strongest biotite alteration in the core (Solomon and Groves, 2000). Disseminated pyrite-chalcopyrite mineralisation and stockwork quartz-pyrite-chalcopyrite veining extend over a 500m by 300m area (Denaro et al., 2007). The Coalstoun deposit and other deposits of this type appear to be derived from aqueous fluids, exsolved from generally porphyritic, intermediate to felsic, granitoid plutonic complexes, (multiple intrusions of diorites, tonalites and granodiorites) showing evidence of crystal fractionation. The magmatism is generally coeval with subduction, and the Cu and Au are derived from magmatic fluids (Ashley et al., 1978).

Several contact metasomatic Cu orebodies are associated with skarns where calcareous sediments of the Yarrol forearc basin were intruded by Triassic granitoids. Most production has been from the Many Peaks and Glassford Creek $\text{Cu}\pm\text{Zn}\pm\text{Au}$ skarns (Fig. 68). Other skarn type deposits are the Biggenden Au-Bi-magnetite deposit and the Ban Ban Zn lode (Murray, 1990; Ashley et al., 1996). The Glassford Creek Cu-Au skarn is located adjacent to mafic granitoids and displays oxidised mineralised assemblages (andradite, magnetite, pyrite, chalcopyrite, bornite) with associated minor Bi, W and Mo minerals. Endoskarn alteration of the intrusives occurs locally (Ashley et al., 1996). The Ban Ban Zn deposit, which crops out as a series of near-vertical skarn lenses within the Biggenden beds, lies adjacent to leucocratic, greisen-altered granite. Replacement of marble has led to the formation of stratabound, sphalerite-rich garnet-dominated skarn, with minor Cu, Pb, Sn, Bi and Ag (Ashley, 1990; Murray, 1990). Ashley and Plimer (1988) support a leucogranite-related hydrothermal origin for the skarn.

At the Biggenden magnetite (Cu-Bi-Au) deposit, a significant source of magnetite for coal washing, magnetite is associated with secondary, massive, crystalline calcite in a skarn developed in thermally altered fine sediments, limestone and andesitic volcanics of the Gympie Group, close to the contact with the Triassic Degilbo Granite (Ashley et al., 1996; Edraki and Ashley, 1999). The overall body is pipelike and seven main magnetite lenses are confined to a 350m by 60m zone (Denaro et al., 2007). Magnetite-garnet-calcite skarn with associated pegmatitic calcite masses have replaced marble and metasedimentary

and metavolcanic hornfelses. Sulphide-rich veins, patches and disseminations are dominated by pyrite and chalcopyrite, but are locally rich in Bi minerals, with minor molybdenite, arsenopyrite and cobaltite. Sulphides are associated with retrogression of the prograde skarn assemblages. Magnetite ore formed in a prograde stage of alteration and the sulphide minerals are part of the retrograde assemblage (Edraki and Ashley, 1999).

3.5.11. Middle to Late Triassic (245-200 Ma) epithermal vein systems, Gympie gold province, northern New England Orogen

The Gympie gold province, located 100-300 km north of Brisbane (Fig. 68), is the most significant mineral province in the New England Orogen. In addition to the Gympie goldfield, this province contains a number of smaller vein and breccia Au deposits, including Mount Rawden and North Arm. These deposits appear to be in part low sulphidation epithermal deposits, and limited data suggest that they overlap in time, in space and, probably, genetically with the porphyry and related Cu±Au±Mo deposits described in section 3.5.10.

Discovered in 1867 and in continuous production for 60 years to 1927, the Gympie goldfield is the sixth largest historical Au producer in Australia. A total of 125 t Au at an average grade of 29 g/t were produced up until 1927. Since 1995 total production has been 8.27 t and current resources and reserves stand at 24.4 t and 6.16 t (www.smedg.org.au/shadcunneenab.htm; accessed December 2008). The Gympie goldfield covers an area of 4 km by 10 km and consists of an extensive, epithermal quartz vein system hosted within the Permo-Triassic mafic to intermediate island arc volcanics and sediments of the Gympie Group (Cranfield et al., 1997). The Gympie Group sits on a Devonian basement of deformed, deep marine, basalt, chert and sediments called the Amamoor Beds. These have been intruded by mid- to late-Triassic granite and diorite. Throughout the region there is a strong spatial and genetic link between base and precious metal mineralisation, Late Permian to Late Triassic plutons and N and NW structural trends (Cunneen, 1996). This mineralisation is mainly epigenetic and is dominantly of Middle to Late Triassic age (Cranfield et al., 1997; Draper, 1998).

All primary mineralisation within the Gympie goldfield consists of quartz calcite veins with free gold and is of two styles, Gympie veins and Inglewood veins (Kitch and Murphy, 1990; Cunneen, 1996; Cranfield et al., 1997). Gympie veins consist of an array fissure filled reefs, 0.1 to 5m wide, that are parallel to the stratigraphy and dip 30-80° to the west. The Gympie veins were the source of most of the Au produced in the Gympie goldfield and have yielded over 100 tonnes of Au since their discovery in 1867 (Kitch and Murphy, 1990). About 70 of these reefs occur in a northwest-trending zone up to 3km wide and 10km long. Higher grade Au mineralisation occurs at the intersection of Gympie veins with beds of carbonaceous shale or siltstone within the Rammutt Formation (Cranfield, 1999). The ore is characterised by free gold, with very small amounts of Cu and As. Pyrite, galena, sphalerite and minor chalcopyrite and arsenopyrite are the common sulphide minerals (Cranfield, 1999). In contrast, the Inglewood Veins strike NW and are sub-vertical. Mineralisation occurs in tabular quartz reefs in large strike-slip fault structures (e.g., the Inglewood Structure) and in strong association with diorite and dolerite dykes (Cunneen, 1996). The Gympie Veins and the Inglewood lode developed contemporaneously, the 'Inglewood Fault vein system' being the feeder and controlling structure in the southern part of the goldfield (Cunneen, 1996).

Oxygen isotope data for quartz are consistent with either a magmatic or metamorphic source (Golding et al., 1987). The Gympie Veins and the associated mineralisation are syntectonic with the cleavage-forming deformation (part of Hunter-Bowen Orogeny), and age dating by Gympie Eldorado Mines Pty Ltd gave 245 ± 14 Ma (method not described) for hydrothermal sericite in andesite adjacent to the Inglewood Fault (Cunneen, 1996). The source of the hydrothermal fluids forming the Gympie Goldfield is considered to be a buried intrusion, defined by airborne magnetics, located immediately southwest of the goldfield (Cunneen, 1996). Geophysical data suggests continuity of this body SE to the Woodnum Igneous Complex south of Gympie which is associated with Mo, Bi, Au and As occurrences and anomalies (P Blevin, pers comm., 2009).

Epithermal Au-Ag precious metal mineralisation with low base metal values also occurs at North Arm (Fig. 68). The North Arm Au-Ag prospect represents a shallow adularia-sericite-type epithermal (Ashley and Andrew, 1992) quartz vein and breccia deposit in Middle to Late Triassic rhyolitic and dacitic volcanic rocks of the North Arm Volcanics (Ashley and Dickie, 1987). At the North Arm Prospect, precious metal mineralisation occurs in veins and stockworks, hydrothermal breccia infillings and as low grade disseminations in altered wallrock. Ashley and Dickie (1987) and Ashley (1987a, b) reported zoned phyllic, argillic and propylitic and K-feldspar alteration in felsic to intermediate volcanic rocks associated with the mineralisation. At nearby Mount Ninderry, the felsic rocks are more intensely altered, with high-level advanced argillic alteration assemblages, but only weak indications of precious and volatile metal enrichment (Ashley and Andrew, 1992). The Late Triassic alteration zone at Mount Ninderry is interpreted to represent an acid sulphate cap over a potential boiling zone containing low-sulphidation style epithermal mineralisation (Ashley et al., 1996). Ashley and Andrew (1992) interpreted that Triassic meteoric waters were dominantly responsible for alteration and mineralisation at North Arm and Mount Ninderry, with the metals sourced from leaching of the rock sequences (Ashley et al., 1996).

Significant Au-Ag production has also come from open cut mining of a subvolcanic breccia system at Mount Rawdon (26.4 Mt at 1.15 g/t Au (29.8 t) and 4.4 g/t Ag (116 t); Angus, 1996; Denaro et al., 2007), which lies in the eastern Gympie Province (Fig. 68). The geology and mineralisation at Mount Rawdon has been described in more detail by Mustard (1986), Brooker and Jaireth (1995), Cayzer and Leckie (1987), Gallo et al. (1990) and Angus (1996). Gold-silver mineralisation is hosted by a sequence of interbedded subaerial pyroclastic flow, surge and ashfall deposits of the Aranbanga Volcanic Group, intruded by coeval dacite bodies and dacite, trachyandesite and trachyte (Denaro et al., 2007). Mineralisation lies adjacent to the intersection of the ENE-trending Swindon Fault and a SSE-trending structure parallel to the Yarrol Fault. These fault trends were important in localising structural complexity and/or mineralisation at Mount Rawdon and other prospects in the region (Angus, 1996). Brooker and Jaireth (1995) described Mount Rawdon as a transitional deposit, as it displays both epithermal and porphyry characteristics. Denaro et al. (2007) describe the deposit as a very high level porphyry system that has intruded its own partially mineralised comagmatic volcanic pile and is centred immediately SW of the original volcanic vent.

Gold and Ag mineralisation at Mount Rawdon postdates dacitic and trachyandesitic intrusion and occurs in two styles: (1) with pervasive fine-grained disseminated pyrite, and (2) in overprinting fractures and irregular veins containing pyrite, sphalerite and minor chalcopryrite, arsenopyrite and galena (Mustard, 1986). Brooker and Jaireth (1995) identified three stages of mineralisation. Stage 1, volumetrically the most significant, contains pyrite, arsenopyrite, sphalerite, galena, native bismuth, Pb-Bi-Ag sulphosalts, matildite, hessite and gold. The gangue minerals consist of chlorite, Mn calcite, epidote, tremolite-actinolite, and apatite. Stage 2 and 3 assemblages are similar. Gold is present as electrum which forms inclusions within pyrite and is closely associated with sphalerite, Pb-Bi-Ag sulphosalts, hessite and matildite (Brooker and Jaireth, 1995). Gold mineralisation is coincident with a zone of pervasive sericite alteration that has overprinted a more widespread, pervasive zone of pre-mineralisation chlorite-carbonate alteration (Denaro et al., 2007).

According to Mustard (1986), mineralisation cannot be linked to one intrusive phase. Dacite, dacite porphyry and trachyandesite are all mineralised to varying degrees which suggests that mineralisation was a prolonged process or involved a number of events. Permeability appears to be the main control on the distribution of the mineralisation. The link between mineralisation and Mid to Late Triassic magmatism has long been inferred, and some of the intrusive phases appear to be contemporaneous with mineralisation and are a source of ore fluids (Angus, 1996). Fluid inclusions in calcite indicate that mineralisation took place from low- to moderate-salinity fluids (0.2-12 wt% NaCl equiv) which were undergoing boiling with temperatures between 220 and 370°C. A few $\delta^{18}\text{O}$ and δD analyses of altered rocks and the vein carbonate indicate that the ore forming fluids had seen some mixing of magmatic and meteoric waters (Brooker and Jaireth, 1995).

3.5.12. Mount Shamrock Au-Ag mineralisation

The Mount Shamrock-Mount Ophir is an Au-Ag mineralised system of probable Triassic age in southeastern Queensland (Fig. 68). Mineralisation is localised by, and related to, a calc-alkaline igneous complex emplaced into Permian-Triassic Gympie Terrane sedimentary rocks (Siemon et al., 1977; Murray, 1986; Williams, 1991). The local country rocks are Permian black siltstones (Biggenden beds), which were intruded by dolerite sills and underwent deformation and metamorphism prior to the igneous and hydrothermal events responsible for the Au-Ag mineralisation (Williams, 1991). The mineralisation is believed to be related to Late Triassic tectonic and volcanic activity (Murray, 1986; Nash, 1986). An account of the regional metallogeny and deposit geology are given by Murray (1986) and Williams (1991), respectively.

Mineralisation is spatially associated with a central volcanic complex intruded by a composite porphyry stock and is controlled by a NE-trending structure parallel to Late Triassic lineaments. Au-Ag-As-rich, Cu-Mo-poor mineralisation occurs in breccias and veinlet networks within pervasively altered rocks (Williams, 1991). The Au-rich zone at Mount Shamrock contained a complex ore paragenesis including pyrite, pyrrhotite, arsenopyrite, chalcopyrite, sphalerite, bismuthinite, native Bi, native Au and tellurides with quartz, calcite and chlorite gangue. Most gold is hosted in sulphides which occur as cement and fracture-fillings in the breccia and as veins in adjacent country rocks. The main alteration, including that associated with most of the Au mineralisation, is characterised by various secondary assemblages of quartz-sericite±albite±carbonate±tourmaline±chlorite±Fe sulphides (+ minor barite and arsenopyrite) formed at temperatures between 350° and 400°C (Williams, 1991). According to Williams (1991), the character and distribution of the alteration and the Au-Ag-As-rich, base metal poor mineralisation at Mount Shamrock is different from typical Cu-Mo-Au porphyry deposits, and has more in common with tectonometamorphic Au deposits formed at much greater depths.

3.5.13. Kilkivan deposits

In the Kilkivan area in southeast Queensland (Fig. 68), straddling the northern part of the North D'Aguilar Block and the eastern margin of the Esk Trough, a variety of small deposit types are hosted in Palaeozoic metamorphic rocks, Permian to Triassic granites and porphyries, and Triassic andesites and rhyolites (Brooks et al., 1974; Bischoff, 1986; Murray, 1986; Nash, 1986). The Kilkivan area has been a minor producer of precious and base metals since the 1860s, and Au, Cu, Hg, Ag-Pb, Mn, and Co-Ni have been mined. Despite the variety of metals and minerals present, no major deposits have been found (Brooks et al., 1974).

Of the numerous small mineral deposits in the Kilkivan area, the majority are of the epigenetic hydrothermal type, and all mineralisation appears to be genetically related to intrusive events. Five deposit categories are identified by Bischoff (1986): granite related deposits; porphyry stock related deposits; dyke-related deposits; porphyry system deposits and speculative intrusion-related deposits.

Granite-hosted deposits are mineralised veins marginal to, but within the granite pluton. For instance, mineralisation related to the Station Creek Adamellite consists of chalcopyrite, tetrahedrite, galena, bornite, sphalerite which occur in a quartz-calcite vein system enclosed by intense phyllic alteration (e.g., Kabunga Cu-Au and Mount Coora Cu-Pb-Zn mines; Bischoff, 1986). Porphyry-hosted deposits are typically Au-Cu±Ag-bearing vein systems surrounded by thin alteration envelopes, within the porphyry intrusive. The best known deposit in the Black Snake Porphyry is the Shamrock Deposit (Bischoff, 1986). Gold-copper-silver mineralisation is located in numerous thin shear veins which are concentrated in four closely spaced ore zones. The mineralisation occurs with quartz, carbonate, pyrite, chalcopyrite, and magnetite. At the surface, the veins are enclosed in quartz diorite porphyry, pervasively altered to a quartz-carbonate-kaolin assemblage. Porphyry-associated deposits occur within the country rocks, closely adjacent to the porphyry intrusive (Bischoff, 1986).

Dyke-related deposits consist of veins in zones contained within, or marginal and parallel to, fine-grained syenite-quartz syenite dykes. For example, the Rise and Shine deposit is a narrow shear and fissure-vein system hosted in greenstones, with gold present in a quartz-carbonate gangue with associated galena, sphalerite, pyrite and arsenopyrite and anomalous Hg. The Gibraltar Rock alteration zone, an example of a porphyry system deposit, is similar to the peripheral parts of known porphyry Cu systems. At Gibraltar Rock, finely disseminated pyrite and chalcopyrite occur over a large area of phyllically altered Neara Volcanics (Bischoff, 1986).

The Kilkivan area is notable because it was the largest Hg producer in Australia (Murray, 1986). Fissure vein and vein breccia-hosted Hg deposits form a distinct group to the west of the main belt of Au and Cu mineralisation (Brooks et al., 1974; Murray 1990). These ?intrusion-related deposits occur in a variety of host rocks including Palaeozoic phyllite, schist, serpentinite, granites and Middle Triassic andesitic volcanics (Murray, 1986), confined to a NNW-trending zone known as the Kilkivan Hg-Belt. Carbonate (-Hg) veins in the Neara Volcanics are typically discordant and occur as composite fissure veins in zones 0.5-3m wide and are superimposed on, or closely adjacent to, interpreted major zones of structural weakness at depth. Cinnabar and lesser pyrite occur as veinlets and disseminations within carbonate-quartz veins, the latter enclosed by alteration envelopes characterised by pervasive carbonatisation, with lesser sericitisation, argillisation and chloritisation (Brooks et al., 1974; Bischoff, 1986). The cinnabar veins are interpreted to have an epithermal origin, perhaps directly related to hydrothermal fluids from an intrusive at depth or they may represent part of a hydrothermal system where heat from granites or the volcanic pile was responsible for the circulation of dominantly meteoric waters which leached the country rocks and mobilised Hg, Cu, As, Sb and Ag into fractures (Bischoff, 1986). The mineralisation appears to have been controlled by faulting along the eastern margin of the Esk Trough, and is probably related to Middle Triassic volcanism (Brooks et al., 1974; Murray, 1986).

3.5.14. Skarn Sn and related deposits, Doradilla district, Lachlan Orogen

In the Doradilla district, which is located in the northern part of the Wagga Sn belt, Sn is localised in steeply plunging shoots within a calc-silicate unit within the Ordovician Girilambone Group (Burton et al., 2007). This unit hosts four prospects (Doradilla, Midway, East Midway and 3KEL) over a 14 km strike length. A JORC-compliant resource totalling 7.81 Mt grading 0.28% Sn has been estimated for the Midway and 3KEL deposits (www.ytcreources.com [accessed 29 November 2008]). In addition, minor Cu was produced between 1901 and 1920 in the nearby Doradilla Cu prospect (Burton et al., 2007).

The Sn is present mostly as cassiterite (Doradilla) and malayaite (Midway, East Midway and 3KEL) and is associated with pyrite, pyrrhotite, galena, sphalerite, chalcopyrite, arsenopyrite, Bi minerals and stannite. The calc-silicate rocks associated with these prospects are dominated by garnet (andradite and grossular)-clinopyroxene (diopside and hedenbergite) skarn with variable amounts of wollastonite, K-feldspar, quartz, titanite, tremolite, carbonate, vesuvianite, fluorite, phlogopite, actinolite, chlorite and magnetite (Kwak, unpub. data, and Young unpub. data, in Burton et al., 2007; Plimer, 1984).

These prospects are spatially associated with the extremely fractionated and moderately reduced, I-type (Blevin, 2004) Midway Granite and a series of quartz-feldspar dykes. Initially these intrusions and the Sn mineralisation were interpreted as Silurian to Early Devonian as these ages were typical of granites in the region (Burton et al., 2007), however, a galena Pb-Pb model age of ~295 Ma (Carr et al., 1995) suggested that mineralisation and possibly the associated magmatic rocks may be significantly younger. To resolve this issue, Burton et al. (2007) dated the Midway Granite and quartz-feldspar porphyry dykes, which yielded ages of ~235 and ~231 Ma, respectively (Burton et al., 2007). These results indicate that Sn deposits in the Doradilla district are significantly younger than other Sn deposits in the Wagga Sn belt. They more closely align with the ages of granites and related deposits in the New England Orogen, although in detail they are slightly younger (235-231 Ma versus 248-238 Ma). These data indicate that magmatism associated with the New England Orogen extended well to the west (500 km) of the mapped extent of the New England Orogen. In addition to these Sn deposits, recent exploration by YTC Resources has identified possible Avebury-type Ni occurrences in serpentinites to the SE of the known Sn deposits.

3.5.15. Uranium deposits of uncertain age and origin, north Queensland

Numerous small, U-F-Mo prospects and two significant deposits (Maureen and Ben Lomond) occur in the Townsville to north Georgetown region (Fig. 67), associated with felsic volcanics and sediments of the Kennedy Province. According to Morrison and Beams (1995) and Mega Uranium (<http://www.megauranium.com>, accessed December 2008), mineralisation includes U-bearing phosphates, sulphates and molybdates in veins or shear infill, as well as replacement zones, closely associated with unconformities. Although mineralisation (U especially) is most likely ultimately sourced from the Kennedy Province felsic magmatism, it is not clear whether these deposits are Carboniferous-Permian in age, representing hydrothermal mineralisation related to the magmatism (Bain, 1977; Morrison and Beams, 1995) or reflect younger, low-temperature, unconformity-related mineralisation (Wall, 2006). Morrison and Beams (1995) suggested the possibility that the deposits may represent the distal portions of polymetallic Sn-W deposits. Ewers (1997) indicated that the deposits are generally small in size. Current (2008) indicated and inferred resource (Canadian NI43-101 compliant) estimates are 2.88 kt U₃O₈ at Maureen and 4.86 kt U₃O₈ at Ben Lomond (<http://www.megauranium.com>).

3.5.16. Mineral potential

Synthesis and analysis of geological and metallogenic data suggest that the Hunter-Bowen cycle was characterised by four geodynamic systems, in order of decreasing age:

1. Kennedy magmatic system;
2. Connors-Auburn Arc-backarc system;
3. Cracow-Chalmers backarc system; and
4. Hunter-Bowen Orogeny.

Although the Kennedy magmatic system largely intrudes the Thomson and North Queensland Orogens and overlaps the other systems, all of these systems appear to be related to the evolution of the New England Orogen. The Connors-Auburn arc-backarc system appears to continue across the Kanimblan Orogeny from the Kanimblan cycle into the early part of the Hunter-Bowen cycle. With the exception of the Doradilla Sn and Ni deposits, Hunter-Bowen activity is not expressed in the Lachlan Orogen.

3.5.16.1. Kennedy magmatic system

As described in section 1.2.6, magmatism in the Kennedy magmatic province spans the period from 340 to 260 Ma, with an apparent younging in ages from west to east, possibly associated with the eastward retreat of the related subduction zone. The granites are mostly I-type with lesser A- and S-types. The A-type granites are relatively young, mostly between 290 and 275 Ma. Compositionally, the granites are dominantly felsic with lesser intermediate and mafic compositions (section 1.2.6). Despite the strong crustal input into the magmatism, these granites are interpreted to be arc-related, possibly in an extensional backarc position relative to the continental arc to the east in the New England Orogen (section 2.6).

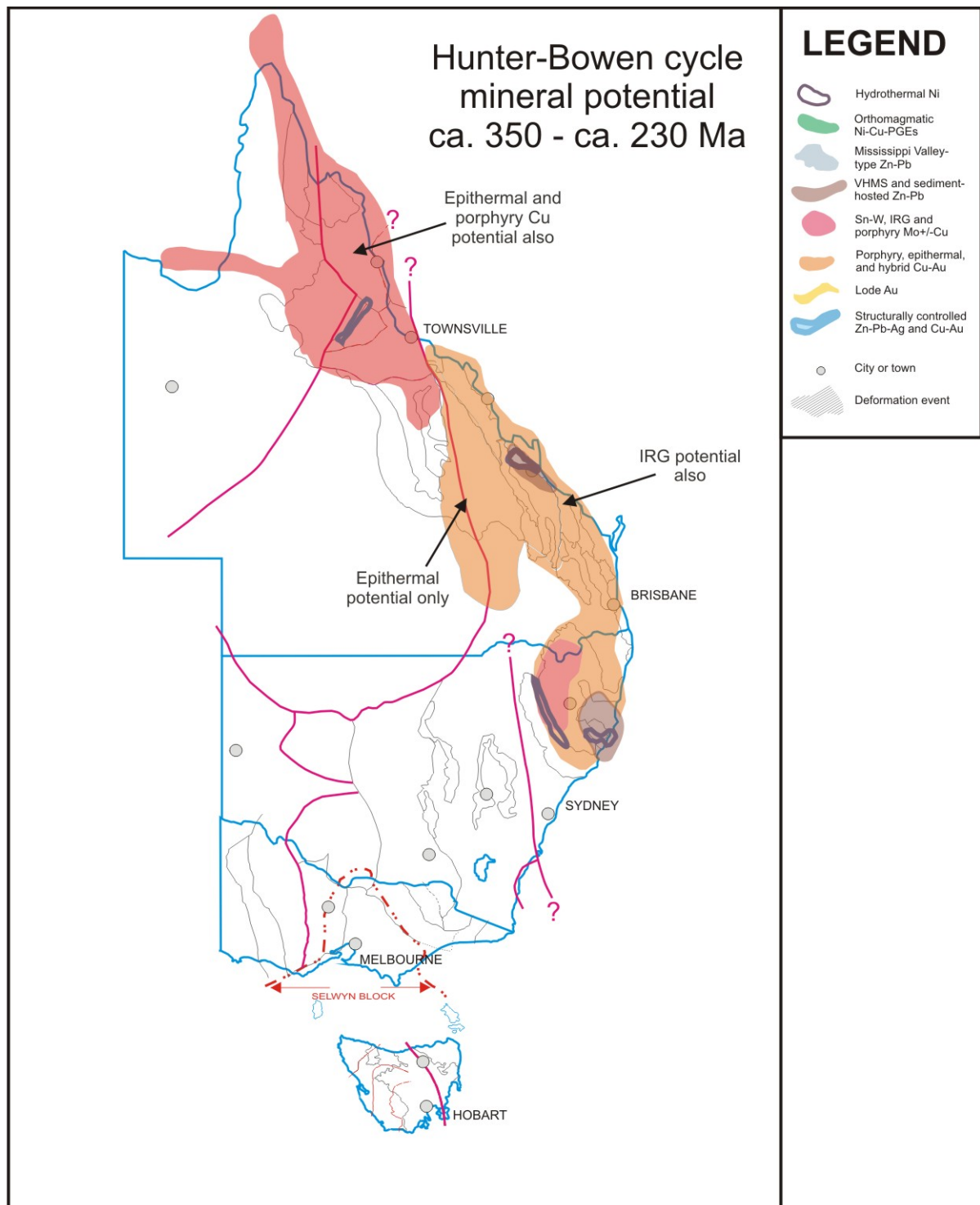


Figure 70. Mineral potential of the Hunter-Bowen cycle.

The Kennedy magmatic province is extensively mineralised, with the age of mineralisation decreasing from west to east like the associated granites. Three styles of intrusion-related deposits are associated with the granites: (1) skarn, greisen and vein-hosted Sn and W deposits; (2) IRG-style vein, skarn-hosted and breccia-pipe hosted Au(Mo-Bi) deposits; and (3) porphyry Cu±Mo±Au and related deposits. Of these three deposit styles, only the Sn-W and Au(Mo-Bi) groups contain significant deposits. The Kennedy magmatic province has significant potential for further discoveries of these two deposit styles (Fig. 70), but much lower potential for the discoveries of porphyry Cu and related deposits as this latter deposit type is more closely associated with oxidised, intermediate intrusions, which are not abundant in the Kennedy magmatic

province. Moreover, these deposits tend to be formed in magmatic arcs, and not in a backarc position as interpreted for the Kennedy magmatic province.

The potential for these deposits may not be uniform across the Kennedy province. Tin-tungsten deposits are largely localised in the east, particularly associated with granites that intrude metasedimentary rocks of the Hodgkinson Province. Intrusion-related gold deposits, in contrast, are more widely distributed. The simplest explanation for this distribution is the oxidation states of the ore-related granites. As demonstrated by Blevin and Chappell (1995), Sn-W mineralisation is related to reduced to strongly reduced granites – their predominance in the Hodgkinson Province suggests that the local sedimentary rocks are intrinsically reducing the magma.

3.5.16.2. Connors-Auburn arc-backarc system

The Connors-Auburn arc was active between 380 and 305 Ma, transgressing the Kanimblan Orogeny ([section 1.3.5](#)) so that it continued from the Kanimblan cycle into the Hunter-Bowen cycle. This arc is defined by subduction-related magmatism with outboard forearc basins of the Yarrol and Tamworth Provinces. As discussed in [section 3.4.3.1](#), no significant mineralisation is known in these areas, possibly because high level deposits were exhumed during later erosion. However, Au-rich low sulphidation epithermal deposits are known to be hosted by Middle to Late Carboniferous (~340 Ma) felsic volcanic rocks in the basal cycle 1 of the Bowen Basin, which is interpreted as a possible back-arc basin to the Connor-Auburn arc ([section 1.3.5](#)). As similar-aged volcanic rocks are also known in the Kennedy magmatic province, for example the Newcastle Range Volcanics, epithermal potential may be more widespread than in the Bowen Basin. Furthermore, subareal or shallow submarine, felsic volcanics of younger age may also have potential for low sulphidation epithermal deposits, and epithermal deposits commonly form in the same metallogenic provinces as porphyry Cu deposits, suggesting that the areas around the Mount Turner and Ruddygore porphyry Cu deposits have epithermal potential.

3.5.16.3. Cracow-Chalmers backarc system

As discussed in [section 1.3.6](#), between 305 and 270 Ma, the New England Orogen was dominated by backarc-related extension associated with slab rollback. This period was characterised by deposition of bimodal volcanic rocks and associated volcanoclastics and siliclastic rocks and by the emplacement of both S-type and I-type granites in the New England Orogen. This extension also initiated the Sydney-Gunnedah Basin system ([section 1.3.6](#)).

Known mineral deposits associated with the Cracow-Chalmers backarc system include the Cracow goldfield and the Mount Chalmers Cu-Au VHMS deposit. The Mount Chalmers deposit is a good example of a potential hybrid VHMS-high sulphidation epithermal Cu-Au system as the deposit is associated with advanced argillic alteration assemblages ([section 3.5.4](#)). The Cracow goldfield is dominated by deposits with characteristics of low sulphidation epithermal deposits ([section 3.5.3](#)). This style of deposit is also known at Mount Terrible in the southern New England Orogen, suggesting that epithermal and related mineral systems may have been widespread through Early Permian volcanic belts in the New England Orogen, possibly with high sulphidation and hybrid deposits favoured toward the east, nearer the inferred offshore arc ([section 2.6](#)). Parts of the Sydney-Gunnedah-Bowen Basin system may also have potential for epithermal-type Au deposits.

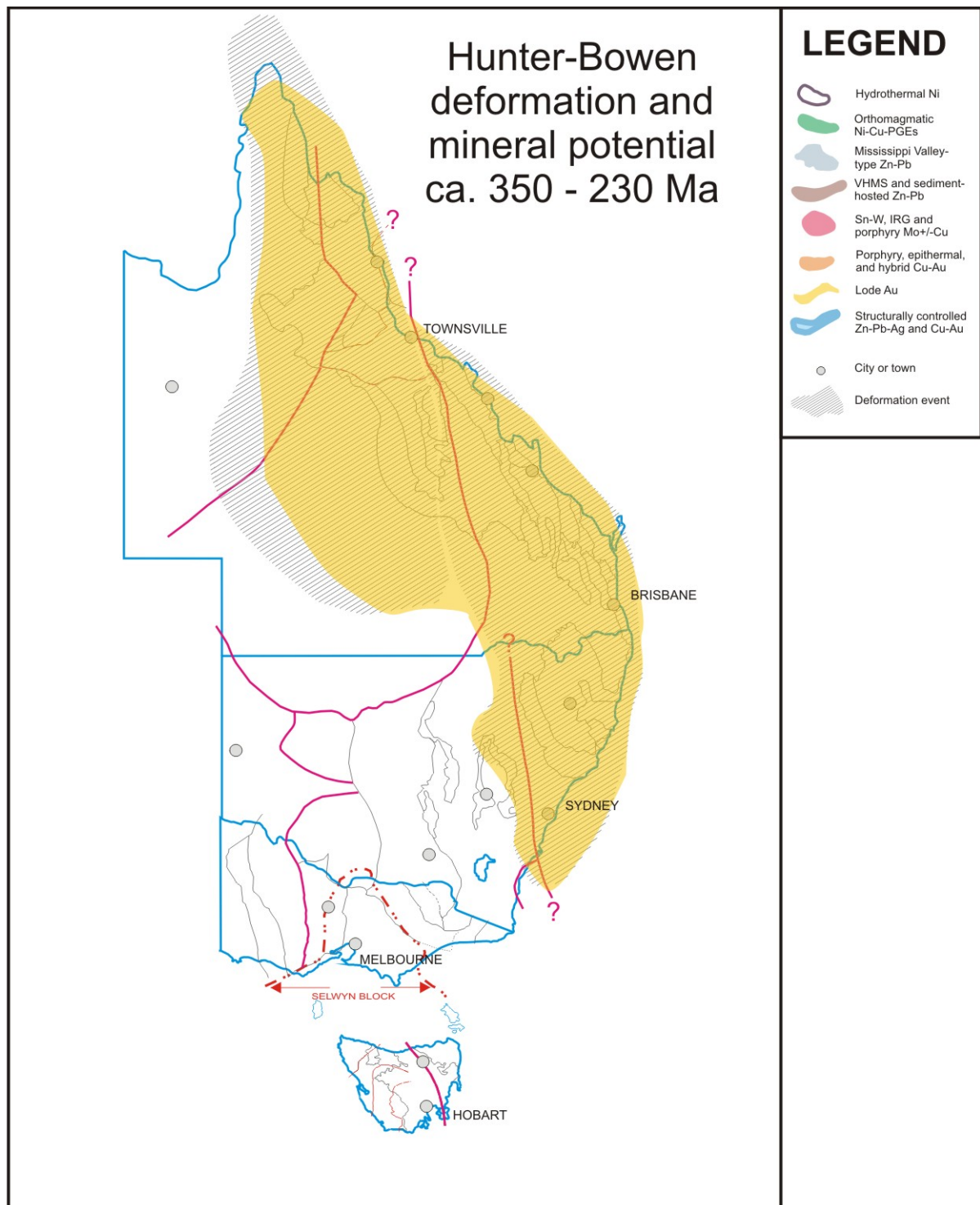


Figure 71. Mineral potential of the Hunter-Bowen Orogeny.

3.5.16.4. Hunter-Bowen Orogeny

At ~265 Ma, the New England Orogen went from extension into contraction when a magmatic arc was re-established on the eastern margin of the Australian continent. As a consequence, a west verging fold-thrust belt developed in the former extensional basins to the west of the developing magmatic arc, and the Sydney-Gunnedah-Bowen system changed from a backarc extensional to a foreland setting, which it retained until the Middle Triassic (Korsch et al., in press; [section 2.7](#)). The Hunter-Bowen Orogeny mostly

affected the New England Orogen and parts of north Queensland, and very minor effects on the Lachlan Orogen.

Metallogenically, one of the earliest effects of this tectonic change was the deposition of lode Au deposits in the Hillgrove Au-Sb district (section 3.5.7) and, possibly, the Hodgkinson goldfield (section 3.5.6). The age (~255 Ma for Hillgrove and Permian for Hodgkinson) combined with the close association with structures suggests that these deposits formed during initial contraction associated with the Hunter-Bowen Orogeny. The Hunter-Bowen Orogeny is developed widely through the New England Orogen, suggesting potential for lode Au deposits through this orogen and its hinterland, and into north Queensland (Fig. 71). This concept is possibly supported by the presence of small lode Au deposits in the Berserker Group near the Mount Chalmers VHMS deposit (Crouch, 1999) in central Queensland. The New England Orogen offers further potential for discovery of lode Au and Au-Sb vein systems in suitable structural settings, perhaps along strike from, or adjacent to, historically productive mines such as at Hillgrove in the southern New England Orogen.

Based on limited age data, granite-related deposits associated with the Hunter-Bowen Orogeny appear to post-date the lode Au deposits, with ages ranging from 250 to 240 Ma. Although the very northern part of the New England Orogen does not contain significant deposits of this age, the central and southern New England Orogen contain an abundance of generally small to medium deposits with diverse metallogenic assemblages. Moreover, there appears to be metallogenic differences between these two areas. The central New England Orogen is dominated by deposits of the porphyry-epithermal system, including the Gympie low sulphidation deposit and a number of sub-economic porphyry Cu-Mo deposits. In contrast, the southern New England Orogen is dominated by granite-related Sn-W deposits related to the New England Batholith, although it also contains the Timbarra IRG deposit.

The New England Orogen has a moderate to high potential for the discovery of additional ore deposits and these will be most likely related to Triassic magmatism. In less eroded portions of the orogen, epithermal and breccia-hosted Au-Ag and porphyry Cu-Au remain prospective at sites of regional structural control; clearly the central New England Orogen retains considerable potential (Ashley et al., 1996). However, a feature of the large number of porphyry Cu-Mo deposits in the New England Orogen is their uniformly subeconomic grade, which may reflect a low S content of the related magmas or possibly an oxidation potential lower than porphyry Cu and porphyry Mo deposits found elsewhere. Another consideration may be the degree of magmatic fractionation, with the Cu-Mo type of deposit being too fractionated to form an economic Cu deposit, but not fractionated enough to form an economic Mo deposit (Solomon and Groves, 2000).

In the southern New England Orogen, the northern part of the New England Batholith contains extensive areas of fractionated, highly prospective leucogranite. However, despite the presence of numerous small mines and mineral deposits, economically viable ore bodies have not been found. Although the Ruby Creek Granite is considered to be the mineralising intrusive phase of the batholith, and potential exists for discovery of new deposits related to cupolas of Ruby Creek and Mole Granite, little potential exists for the discovery of economic deposits (Donchak et al., 2007), with the possible exception of buried intrusions.

The Stanthorpe Supersuite and other Triassic granites have potential for more Timbarra- and other style IRG deposits. At the regional scale, exploration for this style would obviously need to focus on identifying the distribution of the more fractionated, biotite-bearing leucomonzogranite phases of Stanthorpe type plutons and on finding the roof zones of plutons and cupola-style traps for late-stage Au-rich fluids (e.g., see Simmons et al., 1996; Ashley et al., 1996; Mustard, 2001, 2004).

The southern New England Orogen retains considerable potential for occurrence of sheeted vein, disseminated greisen and replacement Sn(W) deposits, but these are likely to be undercover and related to cupolas in the roof zones of buried leucogranites.

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