Geological Society, London, Special Publications

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Geological Society, London, Special Publications 2009; v. 318; p. 329-349
doi:10.1144/SP318.12

Palaeozoic Lachlan orogen, Australia; accretion and construction of continental crust in a marginal ocean setting: isotopic evidence from Cambrian metavolcanic rocks

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Abstract: The Lachlan orogen developed as a classic accretionary orogen in an oceanic setting between the palaeo-Pacific subduction zone and the Australian craton. Direct evidence for the composition and age of the lower crust and the basement to the thick Palaeozoic turbidite fan of the Lachlan orogen is limited. Exposures of Cambrian metavolcanic rocks and geophysical data suggest that most of the basement is the mafic oceanic crust along with possible small fragments of older continental crust. The trace element compositions of Cambrian metavolcanic rocks in the western and central Lachlan orogen are similar to those of volcanic rocks formed in modern back-arc and forearc settings. Pb, Nd and Sr isotopic data from these Cambrian rocks suggest a supra-subduction zone setting with little or no influence of continental crust other than subducted sediment.

Accretionary orogens form from the addition of material to continents from plate margin processes at convergent or transpressive plate boundaries, and are one of the most important 'factories' for generating, recycling and maturing continental crust (e.g. Condie 2007). Palaeozoic–Mesozoic accretionary margins extended over 20 000 km around the margin of Gondwana from the present northern Andes to eastern Australia (Fig. 1a; e.g. Foster & Gray 2000; Cawood 2005). Accretionary processes along the Gondwanan margin added some 20–30% to the area of some continents including Australia, and are still continuing at the edge of the Australian plate in New Zealand (Gray *et al.* 2007).

The Tasmanides (Fig. 1b) of Australia show eastward younging in accretion of submarine fans, fragments of ocean crust, volcanic arcs and forearc basins from the Delamerian–Ross orogen to New Zealand (e.g. Foster & Gray 2000; Cawood 2005; Gray *et al.* 2007). The Lachlan orogen (Fig. 2), which is the central belt within the Australian Tasmanides, is an exceptionally well-preserved accretionary orogen. This belt formed largely in a deep-water oceanic setting from structural thickening and accretion of continental detritus and juvenile igneous components, without the incorporation of extensive blocks of older continental crust, and has great potential for improving our understanding of continental growth and recycling processes (Foster & Gray 2000; Collins 2002*a*; Glen 2005). The Lachlan orogen is widely applicable as a template for circum-Pacific accretionary orogens typified by marginal basins and large volumes of turbidite.

In this paper we review the tectonic history of the Lachlan orogen and present trace element and isotopic data from Cambrian metavolcanic rocks, which form the basement for the orogen. The focus of the geochemical investigation is on the Middle Cambrian calc-alkaline Jamieson–Licola volcanic rocks that formed within the proto-Lachlan marginal ocean basin. The data obtained are used to infer the nature of the lower crust and define the crustal growth and recycling processes in the Lachlan orogen.

The Lachlan orogen

Geological framework

Continental crust of eastern Australia formed along the margin of the supercontinent of Gondwana during the Palaeozoic and early Mesozoic as a result of accretion of oceanic crust, recycled continent-derived turbidite, and volcanic arcs.

From: CAWOOD, P. A. & KRÖNER, A. (eds) *Earth Accretionary Systems in Space and Time*.
The Geological Society, London, Special Publications, **318**, 329–349.
DOI: 10.1144/SP318.12 0305-8719/09/$15.00

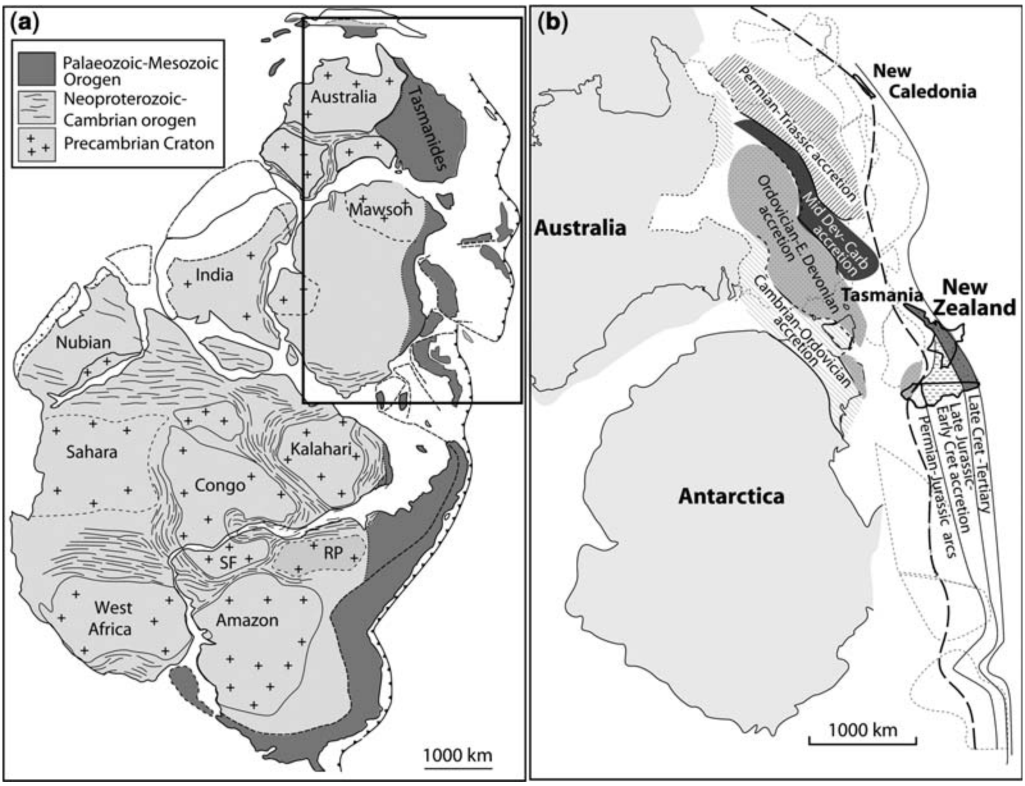

Fig. 1. (**a**) Map of Gondwana showing the location of the Tasmanides within the Palaeozoic–Mesozoic orogenic belts along the margin of Gondwana. (**b**) Reconstruction of Australia and Antarctica along with the timing of accretion for the orogenic belts on the Palaeopacific margin (modified after Gray *et al.* 2007). SF, Sao Francisco; RP, Rio de la Plata.

Eastern Australia is composed of distinct orogenic belts collectively referred to as the Tasmanides (Fig. 2). Accretion occurred with an eastward younging of peak deformation of Middle Cambrian, Ordovician–Devonian and Permian–Triassic age in the respective belts. Outboard of the Tasman orogen, the continental landmass of New Zealand records continuous sedimentation and accretionary prism development above the subduction system from Permian to Late Cretaceous times (Bradshaw 1989; Mortimer 2004).

The Palaeozoic Lachlan orogen (Fig. 2) is a composite orogen dominated by Cambrian to Ordovician turbidites that formed a large submarine fan system comparable in dimension with the Bengal fan (Fergusson & Coney 1992*a*). Rocks of the Western and Central Lachlan orogen are mainly early Palaeozoic quartz-rich sandstone and black shale turbidites, which are laterally extensive over the 800 km width, and have thicknesses upwards of 10 km. The Eastern Lachlan orogen consists of andesite, volcaniclastic rocks, and limestone, as well as quartz-rich turbidite and extensive black shale (VandenBerg & Stewart 1992; Glen *et al.* 2007). Age populations of detrital zircon (U–Pb) and muscovite (^{40}Ar/^{39}Ar) indicate that the turbidites were derived from the Delamerian–Ross and Pan-African orogenic belts throughout Gondwana (e.g. Turner *et al.* 1996; Foster *et al.* 1998; Veevers 2000; Squire *et al.* 2005). The Lachlan turbidite fan accumulated on Cambrian back-arc and forearc crust (Fig. 3a), consisting of predominantly low-K to arc tholeiite basalts and gabbros, high-Mg, low-Ti boninites, and calc-alkaline rocks (Crawford & Keays 1987). The orogen developed by accretion of the oceanic sequences accompanied by marked Late Ordovician–Devonian structural thickening (*c*. 300%) and shortening (*c*. 75%) to form *c*. 35–40 km thick crust (Coney *et al.* 1990; Foster *et al.* 1999; Fergusson 2003). Orogeny included widespread magmatism, which chemically and thermally matured the crustal section.

The Lachlan orogen comprises three thrust systems that constitute the western, central and eastern parts, respectively (Fig. 2; Gray & Foster 1997). The orogen has similar lithotectonic

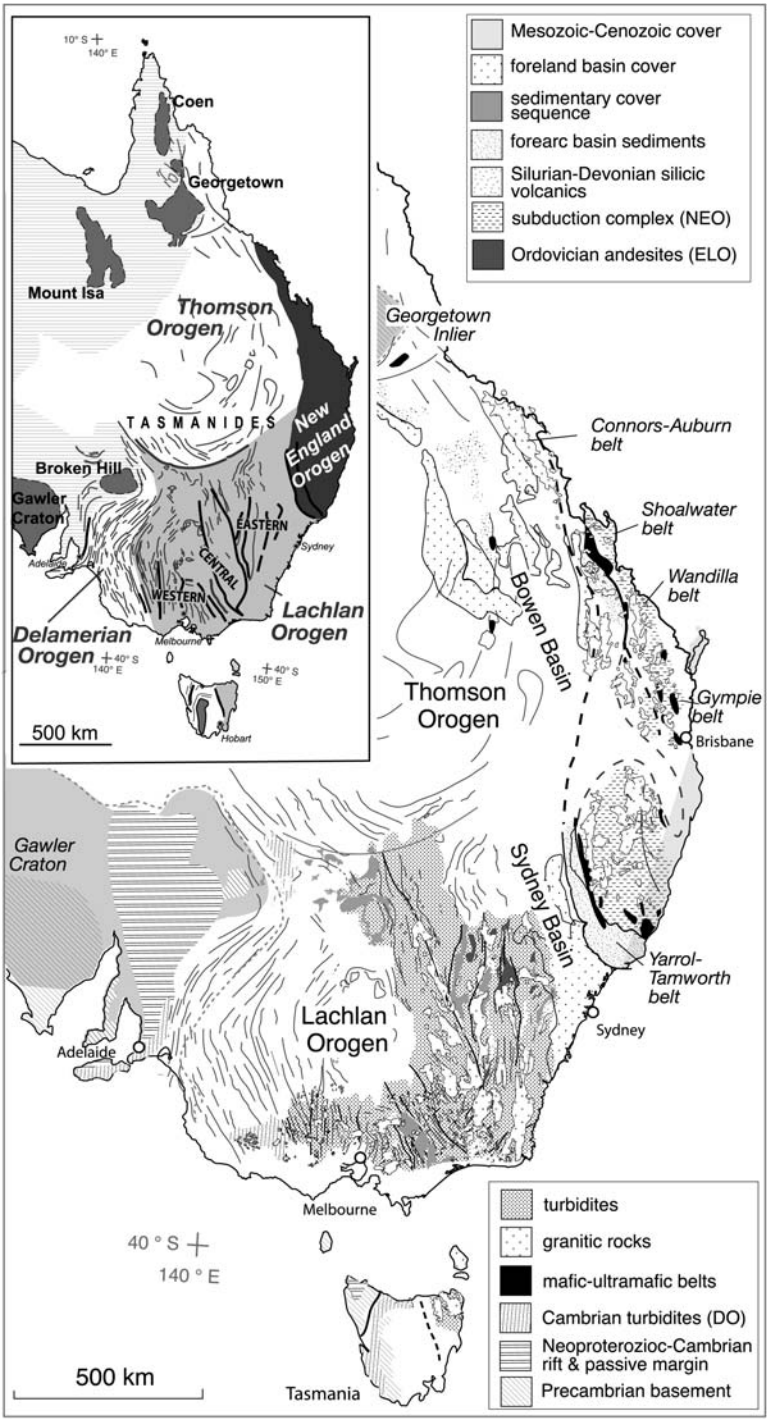

Fig. 2. Geological map of eastern Australia highlighting the main rock units of the Lachlan orogen and the Tasmanides (modified from Gray & Foster 2004). The inset map shows the orogenic belts of the Tasmanides and the three subprovinces (Western, Central and Eastern) of the Lachlan orogen. NEO, New England orogen; ELO, Eastern Lachlan orogen; DO, Delamerian orogen.

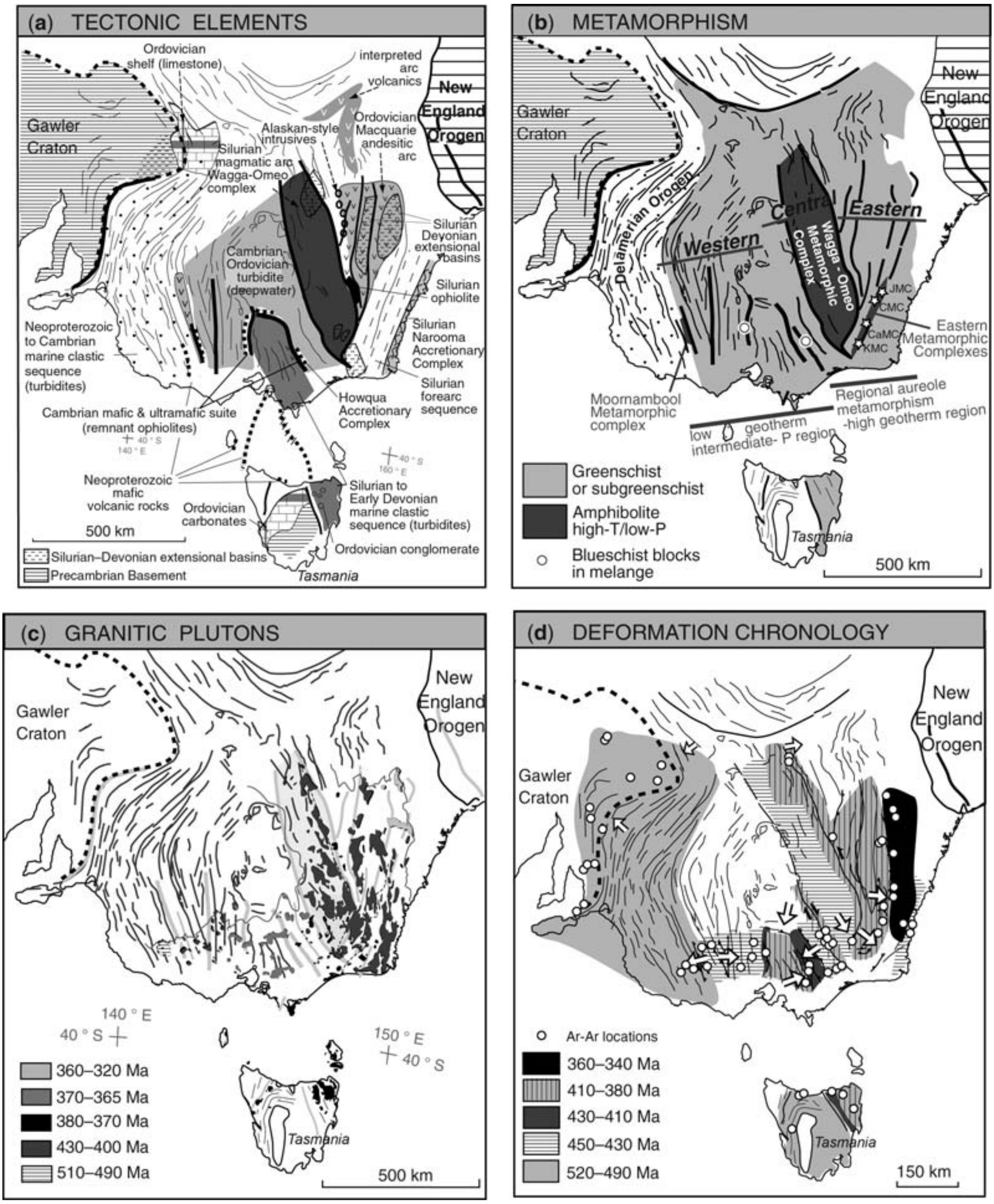

Fig. 3. (**a**) Map showing the major tectonic elements important for understanding the tectonic setting of the Lachlan orogen. The bold lines show locations of major faults and the fine lines show the orientation of the major structural grain. (**b**) Map showing the average grade of metamorphism. (**c**) Map showing the age and location of major granitic plutons. (**d**) Map showing the timing of deformational and metamorphic events across the Lachlan orogen. The arrows show tectonic vergence. JMC, Jerangle Metamorphic Complex; CMC, Cooma Metamorphic Complex; CaMC, Camblong Metamorphic Complex; KMC, Kuark Metamorphic Complex.

assemblages, general structural style and average level of exposure and metamorphism along the entire >1000 km exposed length and across the 700–800 km width of the belt (Fig. 3; Powell 1983; Coney *et al.* 1990; Fergusson & Coney 1992*b*; Gray & Foster 1997, 2004; Foster & Gray 2000; Glen 2005). The Western Lachlan is an east-vergent thrust system with zones of NW–SE- and north–south-trending structures. The Central Lachlan is dominated by NW–SE-trending structures and consists of a SW-vergent thrust belt linked to an extensive high-T/low-P metamorphic complex. The Eastern Lachlan has a north–south structural grain with east-directed faults, with an older, subduction–accretion complex in the southeastern part (Narooma complex; Fig. 3a).

The structure of the Western Lachlan is characterized by chevron folds cut by a series of linked thrust fault systems (Gray & Foster 1998; Korsch *et al.* 2008; Figs 4 and 5). Major faults in the Western Lachlan are spaced about 100–120 km apart, and are polydeformed zones characterized by transposition layering, crenulation cleavages, high, non-coaxial strains, intense mica fabrics, and isoclinal folds (e.g. Gray & Willman 1991; Gray & Foster 1998). The large fault zones contain dismembered Cambrian ophiolite slivers, as well as relicts of serpentinite and mud-matrix mélange incorporating blueschist knockers (Spaggiari *et al.* 2003*a*, 2004).

The Eastern Lachlan orogen is made up of several major, west-dipping fault zones that penetrate to the base of the crust (Glen *et al.* 2002; Gray *et al.* 2006*b*). These faults are crustal-scale imbrications of the Ordovician Macquarie Volcanic Arc and Silurian–Early Devonian platform and deep basin sedimentary rocks. The major faults are spaced 10–15 km apart and have continuous 100–150 km length segments in profile. These major west-dipping faults evolved from Silurian–Devonian normal faults that were reactivated as thrusts during the Silurian–late Early Devonian and the Carboniferous (Glen 1992; Glen *et al.* 1992, 2002). As a consequence, former extensional basins (e.g. Cowra and Hill End basins) have west-directed thrust faults defining their western margins.

Relicts of accretionary complexes are exposed in the southwestern part of the Central Lachlan (Howqua) and the southeastern part of the Eastern Lachlan (Narooma) (Fig. 3a). In the Narooma Accretionary Complex large-scale imbrication is associated with chaotic block-in-matrix mélange, broken formation along high-strain zones, early bedding-parallel cleavage, recumbent folds in turbidites, and structural complexity in cherts. This succession has been interpreted as the outer-arc slope and imbricated zone of an accretionary wedge that was part of a Late Ordovician–Silurian subduction zone (e.g. Powell 1983; Miller & Gray 1997; Offler *et al.* 1998*b*; Fergusson & Frikken 2003). In the Howqua Accretionary Complex an imbricated and chevron-folded, Late Ordovician–Silurian turbidite succession contains mud-matrix and serpentinite-matrix mélanges in the frontal fault zone (Spaggiari *et al.* 2002*a*, 2003*b*, 2004) as well as mud-matrix mélange along major faults within the wedge (Watson & Gray 2001) and a possible detached seamount (Spaggiari *et al.* 2004). These have been interpreted as part of a Late Ordovician–Silurian subduction accretionary wedge (e.g. Foster & Gray 2000; Fergusson 2003).

Deformation and metamorphism started in all of the subprovinces of the Lachlan orogen between *c.* 455 and 440 Ma (Fig. 3d; e.g. Foster *et al.* 1999; Foster & Gray 2007). Metamorphism in the high-temperature–low-pressure Wagga–Omeo Metamorphic Complex and in the eastern metamorphic complexes (Fig. 3b and d) occurred at *c.* 430 Ma (Maas *et al.* 2001; Williams 2001). The interval *c.* 420–410 Ma was characterized by fault reactivation in the Western Lachlan, strike-slip motion on the large boundary faults and mylonitic shear zones of the Wagga–Omeo Metamorphic Complex, and significant exhumation (e.g. Foster *et al.* 1999). The eastern part of the Western Lachlan and the Central Lachlan provinces underwent significant deformation between *c.* 400 and 385 Ma when they collided. The inland parts of the Eastern Lachlan are dominated by *c.* 400–380 Ma deformation with a central and northeastern region of *c.* 380–360 Ma deformation (Glen *et al.* 1992; Foster *et al.* 1999; Fig. 3d). Widespread Silurian extension and basin formation predated this phase of shortening in the Macquarie Arc. These events also overprinted the earlier fabrics in the Eastern Lachlan such as the *c.* 445–440 Ma fabrics in the Narooma Accretionary Complex. Although very widespread, intense Carboniferous deformation is focused in the northern part of the Lachlan orogen in the Eastern subprovince (Powell 1983; Glen 1992).

Silurian–Devonian granitoids make up about 20% of the outcrop area of the Lachlan orogen and up to 36% in the Eastern and Central subprovinces (Fig. 3c; e.g. Chappell *et al.* 1988; Gray & Foster 2004). Most of the plutons were emplaced at pressures ≤ 2 kbar within low-grade upper crustal rocks. Most plutons are post-tectonic and unmetamorphosed, although the older intrusions, such as those within the Wagga–Omeo Metamorphic Complex and Kosciusko Batholith, are significantly foliated and were emplaced synkinematically at mid-crustal depths (e.g. Hine *et al.* 1978; Morand & Gray 1991). Felsic volcanic sequences are also widespread, and form *c.* 15% of outcrop area in the Eastern Lachlan, large post-tectonic caldera complexes in the Western Lachlan, and other fields associated with shallow-level plutons. Older volcanic provinces ($\geq$440 Ma) include the basaltic–andesitic Ordovician Macquarie Volcanic Arc in the Eastern Lachlan orogen (e.g. Glen *et al.* 2007).

The nature of the lower crust beneath the Lachlan turbidite succession is a subject of

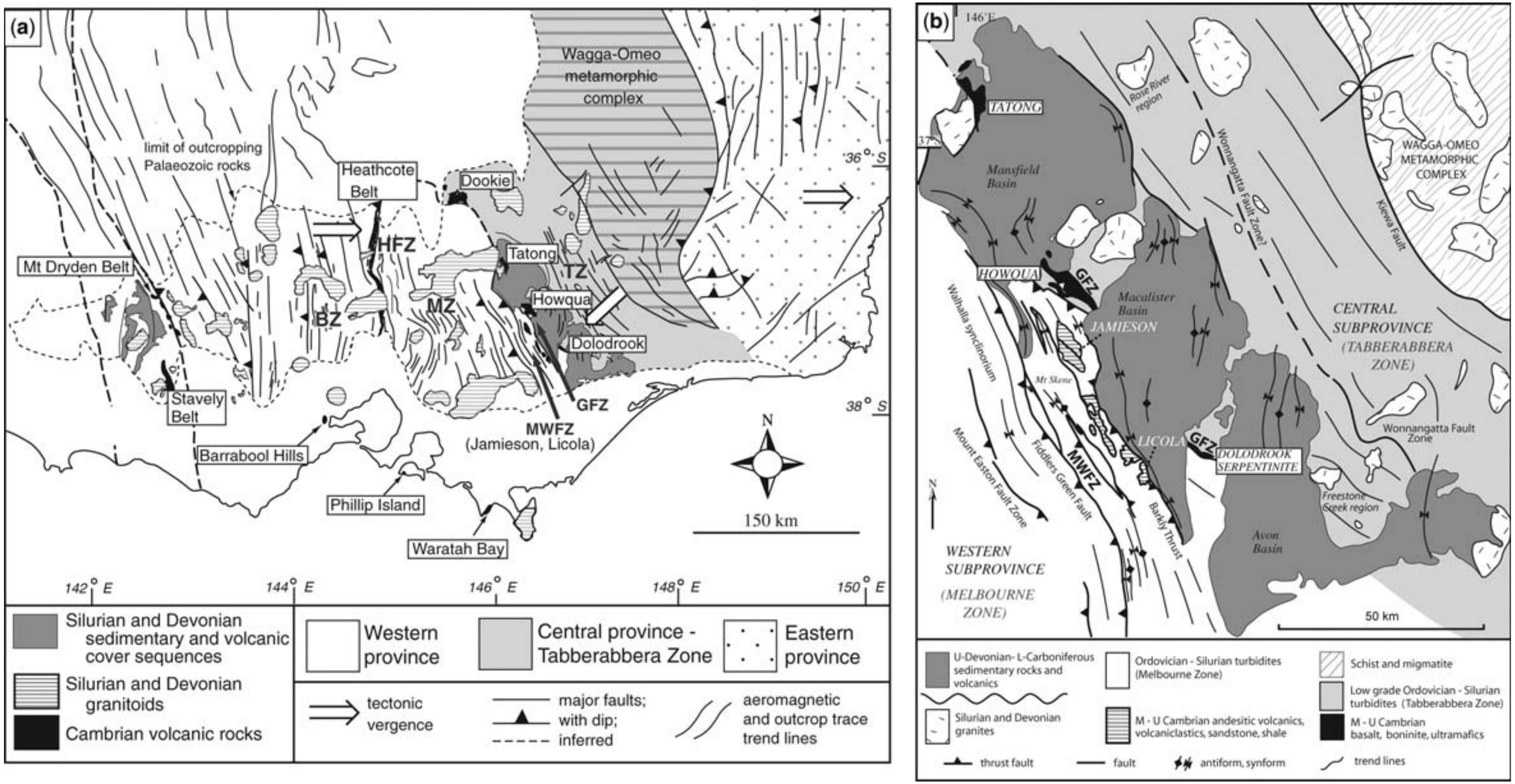

Fig. 4. (**a**) Geological and structural form map of the southern Lachlan orogen highlighting the locations of Cambrian metavolcanic rocks. The Cambrian metavolcanic rocks within the Mount Wellington Belt comprise the exposures within the Governor Fault Zone (GFZ) and the Mount Wellington Fault Zone (MWFZ). BZ, Bendgo Zone; MZ, Melbourne Zone; HFZ, Heathcote Fault Zone. (**b**) More detailed geological map of the boundary zone between the Western and Central subprovinces of the Lachlan orogen. Exposures of the Jamieson–Licola assemblage are within the Mount Wellington Fault Zone (modified from Spaggiari *et al.* 2004).

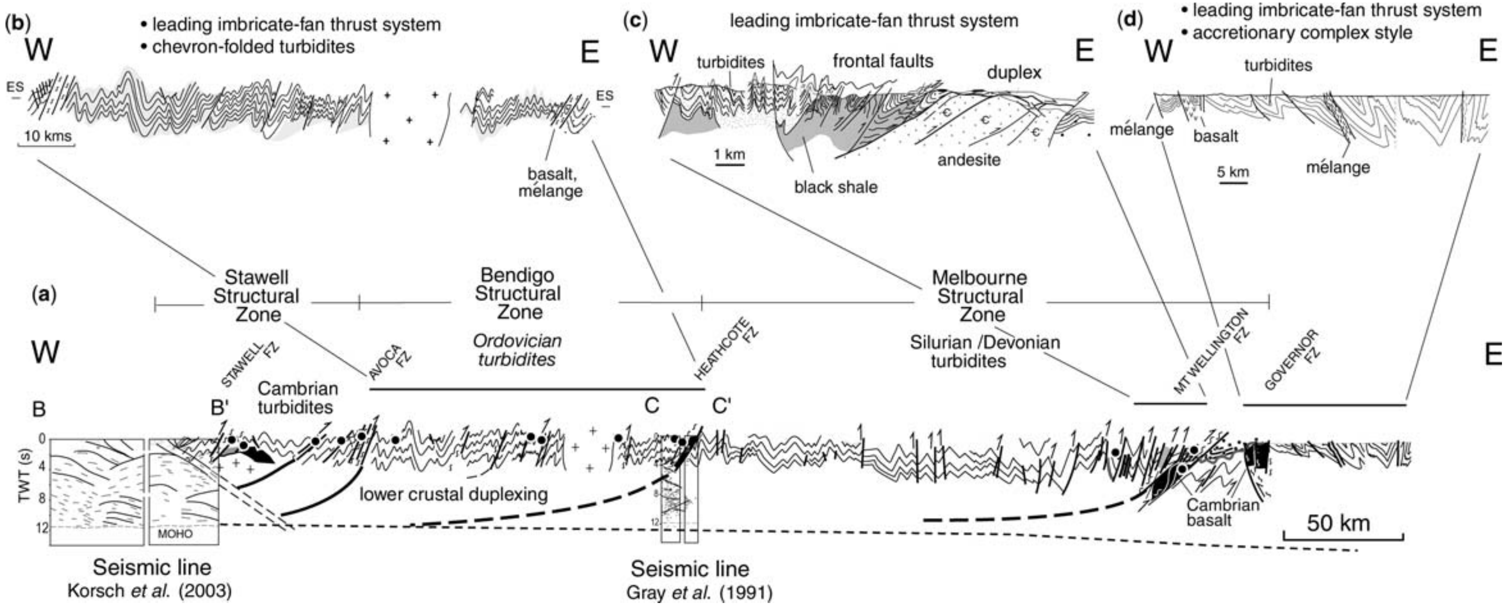

Fig. 5. West to east structural cross-section of the western Lachlan orogen (modified from Gray & Foster 1998; Gray *et al.* 2006*b*).

continued research. In the Western Lachlan Cambrian mafic volcanic rocks of oceanic affinities underlie the quartz-rich turbidite succession (Crawford & Keays 1987), whereas in the Eastern Lachlan the oldest rocks observed are Ordovician volcanic rocks and a Late Cambrian or Early Ordovician chert–turbidite–mafic volcanic assemblage (VandenBerg & Stewart 1992; Glen *et al.* 2007). Chappell *et al.* (1988) suggested that trace element and isotopic data (McCulloch & Chappell 1982) from the Silurian–Devonian granitoids indicated that basement terranes comprising attenuated Proterozoic crust existed beneath the turbidites (see Anderson *et al.* 1998; Handler & Bennett 2001; Maas *et al.* 2001). Whole-rock Nd- and Sr-isotopic data (Gray 1984; Keay *et al.* 1997; Soesoo *et al.* 1997; Collins 1998; Rossiter 2003), Hf- and O-isotopic compositions of igneous zircons (Kemp *et al.* 2005, 2007), similar inherited zircon age populations from S- and I-type granites and the Ordovician turbidite detrital zircon populations (Keay *et al.* 1997; Williams 2001; Kemp *et al.* 2005, 2007), however, indicate that the granitoids are derived from mixtures of the Palaeozoic turbidite, Cambrian oceanic basement and juvenile mantle-derived mafic magmas (e.g. Collins 1998; Kemp *et al.* 2007). Re–Os isotopic data from mantle xenoliths in basalts from the southwestern part of the Lachlan orogen give Proterozoic models ages for the lithospheric mantle (McBride *et al.* 1996; Handler & Bennett 2001). The xenolith locations are within the boundary zone between the Lachlan and Delamerian orogens, and the Delamerian orogen sits on Precambrian continental crust so that these data do not require Precambrian lower crust for the Lachlan orogen.

Seismic reflection profiles in the northeastern Lachlan (Pinchin 1980; Korsch *et al.* 1986, 1993, 1997; Leven *et al.* 1992; Glen *et al.* 1994, 2002; Finlayson *et al.* 2002), and the southern Lachlan (Gray *et al.* 1991; Korsch *et al.* 2002, 2008) have provided an image of the crustal structure (Gray & Foster 2004; Gray *et al.* 2006*b*). These studies show that most major fault zones dip to the west, that steeply dipping faults at the present erosion surface tend to decrease in dip with depth, and that regions between major faults sometimes show complex intersecting networks of both east- and west-dipping faults and shear zones. The reflection studies in the Eastern Lachlan image the structure of the dismembered Ordovician volcanic arc crust (Glen *et al.* 2007). Isostatic considerations require a dense ($>2.9\ \mathrm{g\ cm^{-3}}$) lower crust (e.g. O'Halloran & Rey 1999) matching lower crustal P-wave velocities of $>6\ \mathrm{km\ s^{-1}}$ (Finlayson *et al.* 1979, 1980, 2002; Gibson *et al.* 1981), which are consistent with a mafic oceanic basement for much of the Lachlan orogen.

Many models for the tectonic evolution of the Lachlan orogen suggest development in an intraplate setting involving a single marginal basin floored by attenuated Precambrian continental crust located between a subduction zone and the Australian continent (e.g. Powell 1983; Fergusson & Coney 1992*b*; Li & Powell 2001; Willman *et al.* 2002; Squire & Miller 2003; Braun & Pauselli 2004). More complex scenarios for the evolution of the Lachlan orogen involve multiple subduction systems and microplates in a marginal oceanic setting behind the major, long-lived subduction system (Gray & Foster 1997, 2004; Soesoo *et al.* 1997; Collins 1998; Foster & Gray 2000; Fergusson 2003). The multiple subduction zone–microplate system implies that the basement for the Lachlan orogen is mainly oceanic and not continental, which is consistent with a suggestion made by Crook 40 years ago (Crook 1969). A geodynamic setting involving marginal basin closure by subduction to form the Western and Central subprovinces of the Lachlan orogen, which are within the core of the orogen (Gray & Foster 2004) is suggested by: (1) the presence of dismembered ophiolite slivers along some major fault zones (Spaggiari *et al.* 2003*a*, 2004); (2) the lower temperature, intermediate-pressure metamorphic conditions preserved in metasandstone and slate sequences of the Western Lachlan and external part of the Central Lachlan (Offler *et al.* 1998*a*; Spaggiari *et al.* 2003*b*); (3) the presence of broken formation in the Central and Eastern Lachlan (Miller & Gray 1997; Watson & Gray 2001); and (4) the presence of serpentinite-matrix mélange incorporating blueschist blocks similar to those in the Franciscan Complex of California (Spaggiari *et al.* 2002*a*, *b*, 2003*a*).

Cambrian metavolcanic rocks of the Western and Central Lachlan orogen

Narrow fault-bounded belts of Cambrian ultramafic to andesitic metavolcanic rocks are exposed in the southwestern Lachlan orogen (Crawford 1988). Boninites and low-Ti andesites intruded by arc tholeiites provide evidence for Cambrian back-arc and forearc basin along with island arc settings and indicate a supra-subduction zone origin for the metavolcanic rocks (Crawford *et al.* 1984; Nelson *et al.* 1984; Crawford & Cameron 1985; Crawford & Keays 1987). These belts have, therefore, been interpreted to be faulted slivers of the upper crust of a marginal oceanic basin (Fergusson 2003; Spaggiari *et al.* 2003*a*, 2004; Foster *et al.* 2005). The two largest exposures of metavolcanic rocks are in the Heathcote Belt and Mount Wellington Belt (Fig. 4; Crawford 1988; Spaggiari *et al.* 2004).

The Mount Wellington Belt includes exposed segments of Middle Cambrian, supra-subduction zone ophiolitic rocks at Dookie, Tatong, Howqua, and Dolodrook, which are within the Governor Fault Zone (Fig. 4a and b). The adjacent Mount Wellington Fault Zone contains fault slices of calk-alkaline andesitic rocks of the Jamieson–Licola assemblage that were thrust over the metavolcanic sequences in the Governor Fault Zone (Crawford 1988; Spaggiari *et al.* 2002*b*, 2003*b*; Fig. 5).

The Dookie segment (Fig. 4) is the most northern exposure, and is dominated by tholeiitic basalt, dolerite and gabbro (Crawford 1988). The mafic rocks were metamorphosed at prehnite–pumpellyite to lower greenschist facies and show significant hydrothermal alteration. Epidote, calcite, quartz (± axinite) veins are widespread, and barite veins occur locally. U–Pb zircon data from a hornblende gabbro in the Dookie segment yielded an age of 501 ± 0.7 Ma (Spaggiari *et al.* 2003*a*).

The Howqua segment (Fig. 4) represents the most complete section of ophiolitic rocks in the Mount Wellington Belt (Crawford 1988; Spaggiari *et al.* 2002*b*). The sequence consists of an imbricated section of tholeiitic pillow basalt, hyaloclasite, volcaniclastic rocks, dolerite, and gabbro, in fault contact with and underlain by mafic and ultramafic boninitic lavas and intrusions, underlain by mélange. The mélange contains large fault slivers of both tholeiitic and boninitic rocks, blocks of these rocks metamorphosed to blueschist facies in talc schist matrix, and slivers of Ordovician phyllite, slate and minor sandstone (Spaggiari *et al.* 2002*a*, *b*, 2003*a*, *c*, 2004). The structurally highest, tholeiitic sequence was metamorphosed to prehnite–pumpellyite facies. Metamorphic pressure and temperature, as well as deformation intensity, increase down sequence into the mélange. The metamorphic pattern is interpreted as having formed by accretionary processes during underplating of the mélange, accompanied by duplexing of the upper sequences (Spaggiari *et al.* 2002*a*, 2004). Part of the tholeiitic pillow basalt sequence is conformably overlain by and interbedded with chert and silicified black shale, which, in turn, is conformably overlain by turbidites. Earliest Ordovician (Lancefieldian, La2, *c*. 490 Ma) graptolites occur within the upper part of the chert sequence and Latest Cambrian (Datsonian) conodonts occur approximately midway (VandenBerg & Stewart 1992). These fauna provide a minimum age of 491 Ma for the basalts, suggesting that the basalts are approximately the same age as the Dookie gabbro (*c*. 501 Ma). Tholeiitic rocks show mid-ocean ridge basalt (MORB) to arc tholeiite characteristics and typical Fe-enrichment trends (Crawford & Keays 1987). Older boninitic lavas are also present and most are very mafic (Crawford 1988).

Calk-alkaline rocks of the Jamieson–Licola assemblage are exposed within the Mount Wellington Fault Zone structurally above the Howqua segment (Fig. 4). The sequence is best exposed in the Jamieson River and Licola regions and consists of andesitic to rhyodacitic lavas and minor pyroclastic deposits, voluminous volcanic breccias with interbedded volcanogenic sandstone, siltstone, lenses of black shale, and occasional limestone olistoliths (VandenBerg *et al.* 1995). The lavas are locally pillowed or columnar jointed. The volcanic rocks are strongly deformed, except for some sections of massive lava, and metamorphosed to pumpellyite–actinolite and greenschist facies. Zircons from greenschist-facies andesite lava in the Licola region gave a U–Pb age of 500 ± 8 Ma (Spaggiari *et al.* 2003*a*). This age is in accord with Earliest Ordovician (Lancefieldian, La2, *c*. 490 Ma) graptolites that occur in a thin, black, pyritic shale lens in fault contact with underlying volcaniclastic rocks (VandenBerg *et al.* 1995).

The Cambrian calc-alkaline igneous rocks in the Mount Wellington Belt were erupted through the basement of the Lachlan orogen and predate the Palaeozoic turbidite blanket. Cayley *et al.* (2002) argued that this belt of andesitic to rhyodacitic rocks was erupted through Precambrian continental crust of the Selwyn block because of the presence of intermediate to felsic compositions (see also Scheibner & Veevers 2000; VandenBerg *et al.* 2000). Crawford *et al.* (2003) noted that the compositions of these rocks were very similar to that of the Mount Read volcanic assemblage of Tasmania (Crawford & Berry 1992). Crawford & Berry (1992) interpreted the Mount Read volcanic rocks to be a post-collisional assemblage associated with extension. The volcanic arc-like characteristics of the andesitic rocks, however, are also similar to 'mature-stage', supra-subduction zone rocks described by Shervais (2001).

In the next section we present trace element and whole-rock isotopic data from the metaigneous rocks in the Mount Wellington Belt along with a smaller number of samples from the Heathcote Belt in Tables 1 and 2, and Figures 6–9. The trace element, rare earth element (REE) and isotopic data for the metavolcanic rocks of the Jamieson–Licola assemblage and related metavolcanic rocks (Tables 1 and 2) supplement published results (e.g. Nelson *et al.* 1984; Crawford 1988; VandenBerg *et al.* 1995). These data have significant implications for the Cambrian tectonic setting of the Lachlan marginal basin and the nature of the lower crust, because they are not contaminated by continent-derived turbidites such as the Ordovician–Devonian igneous rocks of the Lachlan orogen.

Table 1. *Elemental and isotopic data from Cambrian metavolcanic rocks in the Western Lachlan orogen*

Exposure area/fault slice		Dookie			Jamieson					
Sample number:	CS0040	CS0045	CS0046	CS9953	CS0054	HED5-TRAY69	HED5-TRAY70	320b	328a	1064a
Rock type:	gabbro	basalt	basalt	andesite	andesite	basalt	andesite	rhyolite	andesite	andesite
Major elements (wt%)										
SiO_2	49.2	48.0	48.1	54.1	54.4	49.2	57.4	73.4	54.1	56.4
Al_2O_3	16.0	17.0	16.1	16.2	14.4	14.9	14.8	10.8	16.6	19.6
TiO_2	1.0	0.4	1.2	0.4	0.3	0.3	0.4	0.4	0.4	0.3
Fe_2O_3	11.9	12.8	13.2	8.1	10.0	7.7	7.6	7.3	7.4	6.6
MgO	7.6	7.3	7.1	4.8	7.4	5.7	5.5	2.0	4.6	3.2
MnO	0.2	0.3	0.2	0.1	0.1	0.1	0.1	0.0	0.1	0.1
CaO	8.2	7.5	9.1	11.4	7.0	3.3	4.0	0.4	8.2	8.1
Na_2O	4.0	3.9	3.2	0.1	2.8	1.5	3.1	0.0	3.5	3.8
K_2O	0.1	0.2	0.1	0.1	1.0	1.8	0.9	2.2	1.0	1.1
P_2O_5	0.1	0.1	0.1	0.1	0.2	0.1	0.1	0.5	0.2	0.2
LOI	2.6	2.1	1.9	4.5	2.4	6.2	6.4	2.9	2.1	2.0
Total	100.8	99.6	100.5	99.8	99.8	99.7	100.4	99.9	100.4	100.7
Trace elements (ppm)										
Sc	43.0	44.1	42.1	24.4	28.0	22.8	23.0	14.0	21.0	23.2
V	238.1	301.6	319.8	196.1	201.9	196.2	186.4	71.7	210.9	206.5
Cr	15.0	276.5	170.1	192.5	292.6	69.7	103.9	110.1	172.4	48.8
Co	47.0	49.4	46.9	21.9	32.7	17.8	21.1	10.4	21.6	13.9
Ni	67.2	107.3	90.6	27.7	46.1	13.1	16.0	35.7	38.8	18.4
Cu	199.6	159.8	173.4	132.8	123.0	170.2	196.1	69.8	138.2	57.6
Zn	70.2	91.6	87.1	59.0	68.3	130.9	79.9	76.7	73.6	42.7
Ga	11.6	15.5	16.6	20.4	14.9	15.1	15.6	14.0	18.1	17.7
Rb	1.0	3.6	2.2	1.6	16.6	30.4	14.9	45.6	15.9	27.2
Sr	68.6	124.9	274.1	185.2	389.3	110.6	116.8	14.4	626.2	418.9
Y	17.3	26.3	26.8	9.6	14.7	9.2	9.5	41.7	16.4	12.9
Zr	43.7	63.1	67.4	68.7	71.1	83.8	94.7	85.7	107.7	74.2
Nb	2.7	3.9	4.2	2.9	2.8	3.7	4.1	8.4	6.1	4.4
Ba	32.2	5060.5	567.6	39.0	1606.0	657.8	313.0	1516.9	513.7	665.9
La	2.2	3.7	3.8	7.9	11.8	7.4	11.4	19.5	32.6	13.6
Ce	6.1	9.5	10.2	16.1	23.6	15.2	22.6	31.7	43.3	24.6
Pr	1.0	1.5	1.6	2.1	2.9	1.8	2.6	4.4	6.9	2.9
Nd	5.1	8.0	8.4	9.2	13.5	7.8	11.1	20.0	29.6	12.7
Sm	1.7	2.6	2.7	2.0	2.9	1.7	2.1	4.4	5.2	2.6
Eu	0.6	2.2	1.0	0.6	1.1	0.5	0.6	1.4	1.3	0.8
Gd	2.1	3.5	3.5	1.6	2.6	1.7	1.8	5.0	3.9	2.4
Tb	0.4	0.6	0.7	0.3	0.4	0.3	0.3	0.9	0.5	0.4
Dy	2.8	4.1	4.2	1.5	2.3	1.6	1.6	5.5	3.1	2.1
Ho	0.6	0.9	0.9	0.3	0.5	0.3	0.3	1.2	0.6	0.4
Er	1.6	2.4	2.4	0.8	1.2	0.9	0.9	3.4	1.5	1.1
Tm	0.3	0.4	0.4	0.1	0.2	0.1	0.1	0.6	0.2	0.2
Yb	1.7	2.6	2.6	0.8	1.2	1.1	1.0	3.8	1.5	1.1
Lu	0.3	0.4	0.4	0.1	0.2	0.2	0.2	0.6	0.2	0.2
Hf	1.2	1.8	1.9	1.9	2.0	2.4	2.7	2.4	3.2	2.1
Ta	0.2	0.3	0.3	0.2	0.2	0.3	0.3	0.6	0.7	0.4
Pb	0.7	1.1	0.7	7.9	5.7	5.8	3.9	7.9	14.4	9.8
Th	0.2	0.3	0.3	3.5	3.6	4.5	5.5	8.9	10.1	4.7
U	0.1	0.1	0.1	0.9	1.1	1.5	1.7	2.5	2.6	1.4
Isotopes										
Sr (ppm)	65.6	110.3	250.8	185	383.1	101.7	107.1		584.5	380.1
Rb (ppm)		3.3		nd	nd	nd	17.2		20.2	35.1
Sm (ppm)	1.2	1.8	1.6	1.7	2.7	1	1.3		nd	1.6
Nd (ppm)	3.6	5.4	4.9	7.5	12.3	4.4	6.7		23.1	17.5
$^{143}Nd/^{144}Nd$ (corrected)	0.512927	0.51291	0.512915	0.512654	0.512637	0.512649	0.512594		0.512584	0.512628
ε_{Nd} (500 Ma)	4.8	5	5.2	4.1	4	4	4.2		4.4	4.1
$^{87}Sr/^{86}Sr$	0.7048	0.7068	0.7049	0.7044	0.7055	0.7103	0.7079		0.7049	0.7055
$^{87}Sr/^{86}Sr$ (500 Ma)	0.7045	0.7062	0.7048	0.7042	0.7046	0.7048	0.7053		0.7044	0.7042
$^{206}Pb/^{204}Pb$ (500 Ma)		17.879		17.895	17.875	17.502		17.352		
$^{207}Pb/^{204}Pb$ (500 Ma)		15.549		15.528	15.553	15.509		15.547		
$^{208}Pb/^{204}Pb$ (500 Ma)		37.688		37.717	37.702	37.355		37.23		

(*Continued*)

Table 1. *Continued*

Licola			Waratah Bay	Heathcote		Howqua		Phillip Island
CS9761 andesite	C9762 andesite	CS9857 andesite	CS9736 basalt	CS9707 boninite	CS9708 andesite	CS9741 boninite	CS9751 basalt	PI basalt
61.5	57.6	57.9	48.3	56.4	64.6	52.4	50.0	47.7
14.7	18.7	13.5	13.8	9.7	12.7	1.8	13.1	19.4
0.4	0.3	0.4	1.5	0.3	0.3	0.1	1.9	1.0
6.1	5.7	7.7	14.2	9.8	6.3	11.7	14.8	10.3
4.3	3.7	7.0	7.4	10.1	6.1	22.8	6.5	9.1
0.1	0.1	0.1	0.2	0.2	0.1	0.2	0.2	0.2
5.5	7.9	8.0	8.1	6.7	3.8	6.8	7.5	7.7
5.3	3.1	1.8	3.5	4.1	4.0	0.3	3.2	4.2
1.1	0.9	0.8	0.5	0.1	1.8	0.0	0.1	0.2
0.3	0.2	0.2	0.1	0.1	0.1	0.0	0.2	0.1
1.4	2.5	3.0	2.6	4.1	2.2	4.0	3.1	1.3
100.6	100.7	100.4	100.3	101.6	101.4	100.1	100.6	101.0
16.1	15.2	26.9	40.4	28.6	15.5	18.8	37.5	41
139.8	147.5	180.8	306.2	211.5	99.9	76.4	323.6	215
167.5	152.3	259.5	143.9	876.0	364.9	4104.3	80.6	471
24.2	15.9	23.6	45.6	43.5	34.8	77.1	44.8	54
39.5	35.8	47.5	77.1	208.5	70.1	567.3	48.5	144
66.9	87.4	150.6	161.2	29.1	32.5	17.7	153.4	83
57.3	52.0	59.6	85.2	54.5	48.3	61.4	102.7	63
12.6	16.6	13.6	15.3	7.5	11.0	1.2	16.7	12
48.5	37.5	26.1	6.5	3.3	19.7	0.8	1.8	3
350.0	986.9	596.6	198.2	49.9	125.0	49.8	156.8	146
13.5	12.4	13.5	28.9	7.7	8.9	0.7	32.7	20
115.8	111.5	109.2	84.8	28.0	88.1	14.3	109.9	49
7.9	7.3	6.4	5.4	0.6	4.9	0.4	8.3	5.1
797.0	690.7	773.2	170.6	171.1	347.2	16.0	207.3	51
48.6	43.6	20.2	4.6	3.8	7.5	1.0	6.6	3.52
85.5	76.6	38.5	11.7	8.1	16.0	2.3	16.5	8.75
8.7	7.9	4.4	1.8	1.1	2.0	0.3	2.4	1.25
34.2	31.2	18.7	9.6	5.3	8.4	0.8	12.8	5.98
5.2	4.8	3.6	3.0	1.3	1.8	0.1	3.8	1.80
1.2	1.2	1.0	1.0	0.4	0.5	0.0	1.3	0.62
4.3	4.1	2.6	4.0	1.3	1.7	0.1	4.8	2.58
0.6	0.5	0.4	0.7	0.2	0.3	0.0	0.9	0.48
2.4	2.2	2.3	4.6	1.2	1.5	0.1	5.4	3.13
0.4	0.4	0.4	1.0	0.3	0.3	0.0	1.1	0.69
1.3	1.2	1.2	2.7	0.7	0.8	0.1	3.0	1.90
0.2	0.2	0.2	0.4	0.1	0.1	0.0	0.5	0.30
1.2	1.1	1.2	2.9	0.8	0.9	0.1	3.3	2.13
0.2	0.2	0.2	0.5	0.1	0.1	0.0	0.5	0.34
3.4	3.2	3.2	2.3	0.7	2.4	0.2	3.0	1.30
0.6	0.6	0.5	0.4	0.1	0.4	0.0	0.5	0.33
14.7	40.5	8.2	0.6	1.4	6.1	0.9	1.1	2.12
25.3	22.7	9.9	0.5	0.9	1.8	0.5	0.7	0.47
4.8	5.6	3.0	0.2	0.7	0.6	0.2	0.2	0.12
325.7	989.9	548.3	184.9	47.9	146.2	47.9	146.2	139
202.9	70.9	nd	nd	3.3	26.6	0.7	nd	nd
3.6	nd	3.4	2.4	0.9	1.1	nd	2.7	1.1
22.9	21.6	17.5	7.6	3.8	4.9	nd	8.8	3.6
0.512313	0.512322	0.512564	0.512887	0.51277	0.512557		0.512867	0.512827
0.1	0.2	3.5	5.1	5.5	2.4		5.2	4.2
0.7083	0.7065	0.7056	0.7066	0.7088	0.7078	0.7087	0.7055	0.7065
0.7055	0.7058	0.7047	0.7059	0.7074	0.7045		0.7053	0.7060
18.333	18.222	17.892		18.482			18.732	18.586
15.588	15.583	15.535		15.534			15.545	15.616
37.282	37.961	37.568		38.025			38.551	38.357

Major elements analysed by X-ray fluorescence. Trace elements analysed by inductively coupled plasma mass spectrometry (ICP-MS) using an Element-2 at the University of Florida. Sr isotopic ratios measured by thermalionization mass spectrometry using a Micromass Sector 54; Nd and Pb isotopic ratios measured by multi-collector ICP-MS using a Nu-Plasma instrument from spiked solutions at the University of Florida. LOI, loss on ignition.

Table 2. *Locations of Cambrian metavolcanic rocks*

Sample number	Rock type	Fault slice	Location	Map coordinates
		Dookie		
CS0040	Gabbro		Kellows Road Quarry	378000E, 5979000N
CS0045	Basalt		Mt. Major, south of communication towers	383800E, 5975400N
CS0046	Basalt		Mt. Major, south of communication towers	384100E, 5975420N
		Jamieson		
CS9953	Andesite		Jamieson River, Wren's Flat	444205E, 5866760N
CS0054	Andesite		South of Mt. Sunday Track, near eastern margin	446705E, 5864740N
TRAY69	Basalt		Hill 800 (drill core)	444936E, 5868966N
TRAY70	Andesite		Hill 800 (drill core)	444936E, 5868966N
320b	Rhyolite		Handford Creek Track, rockfall	443243E, 5863869N
328a	Andesite		Prickle Spur Track	446270E, 5867470N
1064a	Andesite		Jamieson River (Silver Mine Spur Track)	
		Licola		
CS9761	Andesite		Jamieson–Heyfield Road, Wallaby Creek	462365E, 5839120N
CS9862	Andesite		Jamieson–Heyfield Road, Wallaby Creek	463200E, 5835660N
CS9857	Andesite		Jamieson–Heyfield Road, junction of Violet Hill Track	458675E, 5840765N
		Waratah Bay		
CS9736	Basalt		Point Grinder	410000E, 5693750N
		Heathcote		
CS9707	Boninite		Sheoak Gully, Shuran's Lane property	294250E, 5917650N
CS9708	Andesite		Lady's Pass, Colbinabbin turnoff	295600E, 5923000N
		Howqua		
CS9741	Boninite		Cold Creek, trail south of Howqua Track	437800E, 5885570N
CS9751	Basalt		Howqua River, west of Noonan's Hut	443100E, 5883950N
		Phillip Island		
PI	Basalt		Near Watt Point, south coast of Phillip Island	

Map coordinates are from the Australian Map Grid, Zone 55.

Geochemistry of the Cambrian volcanic rocks

Normalized trace element plots show enrichment in large ion lithophile elements (LILE) and strong depletion in high field strength elements (HSFE) for the andesitic rocks (Fig. 6). REE trends (Fig. 7) of the andesitic rocks show light REE (LREE) enrichment and flat heavy REE (HREE) patterns typical of oceanic island arcs lacking deep melting processes in the presence of garnet. The one rhyolitic rock is more enriched in HREE. Eu anomalies are not exhibited by the samples so that plagioclase fraction was limited. All of the basaltic rocks show HSFE compositions similar to MORB, but with arc-type enrichment in LILE consistent with oceanic back-arc basin igneous rocks. Tholeiitic basalts show flat REE patterns, which are also typical of oceanic back-arc basins. The Howqua boninite is extremely depleted in REE. Nelson *et al.* (1984) reported similar trace element data for these rock types.

Initial $^{87}Sr/^{86}Sr$ (500 Ma) ratios for all of the samples analysed range from about 0.704 to 0.708 (Fig. 8). There is significant evidence for seawater alteration and mobilization of Sr isotopes in some samples so that some do not record primary values (see Nelson *et al.* 1984). ε_{Nd} (500 Ma) values for the andesitic rocks are all positive and range from 0.1 to 4.2, and for the basaltic rocks are in the range of *c.* 4–8. The basaltic rocks approach the depleted mantle values for 500 Ma. The boninites also fall into this range. Nelson *et al.* (1984) reported one boninite sample with an initial ε_{Nd} of -9, which is far less radiogenic than anything we measured and is probably an artefact of very low Nd. Common Pb isotopic ratios for almost all of the samples, including the Jamieson–Licola andesites and rhyolite, lie above the curve for Bulk Earth and below that for average continental crust (Fig. 9).

Taken together, the elemental and isotopic data are consistent with previous interpretations of Nelson *et al.* (1984) and Crawford (1988) that the Cambrian metavolcanic rocks in the Western

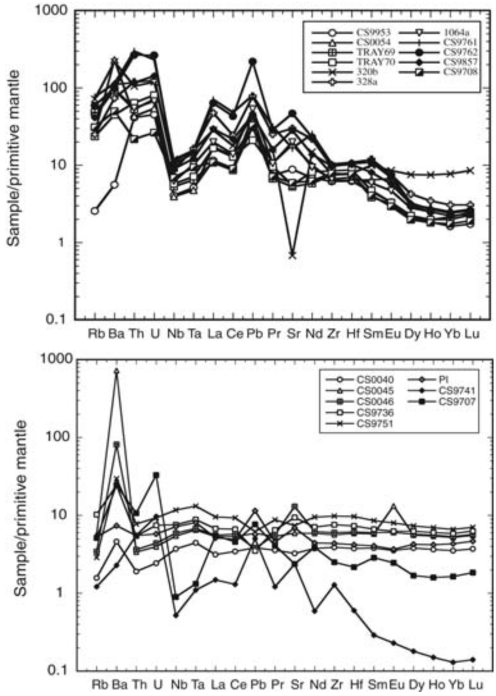

Fig. 6. Trace element variation diagrams for Cambrian metavolcanic rocks from the Lachlan orogen. The top plot includes the andesite samples and the bottom plot shows the samples of basalt and boninite. The sample numbers correspond to those in Table 1.

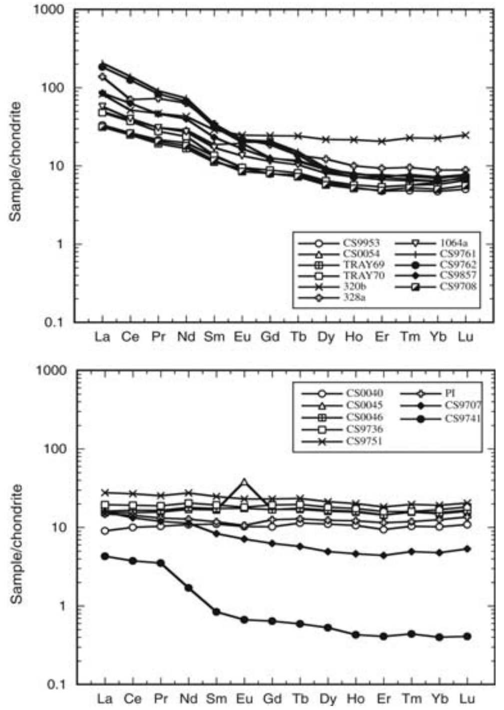

Fig. 7. Rare earth element variation diagrams for Cambrian metavolcanic rocks from the southern Lachlan orogen. The top plot includes the andesitic samples and the bottom plot shows the samples of associated basalt and boninite. The sample numbers correspond to those in Table 1.

and Central Lachlan were formed in a western Pacific-style oceanic marginal basin off the Australian continent. The older boninitic rocks may record early rifting of the forearc of a Delamerian subduction–arc system (Foster *et al.* 2005). The younger tholeiitic rocks resemble basalts erupted during spreading of the proto-Lachlan back-arc basin, whereas the calc-alkaline rocks of the Jamieson–Licola assemblage were probably erupted on the back-arc basin crust in an ocean volcanic arc setting. LILE enrichment and strong depletion in HFSE in these rocks are consistent with derivation from metasomatized mantle above a subduction zone. The isotopic data are also consistent with an oceanic arc and back-arc setting. The spread in ε_{Nd} values to values approaching zero as well as the Pb isotopic data are consistent with sediment contamination in a subduction environment. There is no requirement in the geochemical data that Precambrian crust underlies the Jamieson–Licola volcanic rocks or any other part of the Western or Central Lachlan. The results support the interpretation that the basement for the Lachlan orogen is largely oceanic, and that no volumetrically significant Precambrian continental blocks contributed to the geochemical signatures of these volcanic rocks. Geochemical and isotopic data from the younger Ordovician Macquarie Arc andesites and basalts also indicate construction

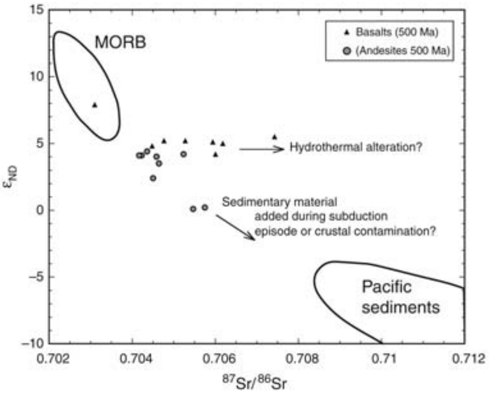

Fig. 8. Plot of ε_{Nd} v. $^{87}Sr/^{86}Sr_i$ for the Cambrian volcanic rocks. The values are corrected to 500 Ma.

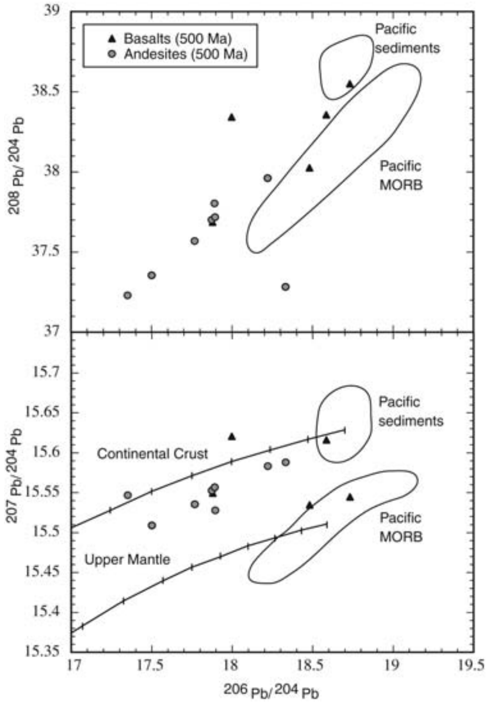

Fig. 9. Plot of common Pb isotopic data for the Cambrian metavolcanic rocks.

on Cambrian oceanic crust (reviewed by Glen *et al.* 2007).

It is possible, however, that the intermediate Pb isotopic values and less radiogenic Nd isotopic values of some samples reflect minor contamination of older continental material. If Proterozoic lower crust is present under the Jamieson–Licola volcanic rocks then the Cambrian volcanic rocks were not significantly contaminated by it, or the Precambrian component was relatively young because it has not imparted an obvious signature on the isotopic data. Direct correlation between the Jamieson–Licola volcanic rocks and those in the Mount Read Belt of Tasmania (e.g. Cayley *et al.* 2002; Crawford *et al.* 2003), are possible only if Precambrian crust of the Tasmanian microcontinent was much thinner under central Victoria. The results suggest that the Jamieson–Licola volcanic rocks developed as a volcanic arc above the Lachlan back-arc basin crust at the same time that the Mount Read volcanic rocks were erupted on the Tasmanian microcontinent, and the Mount Stavely volcanic rocks (Fig. 4) were erupted on the edge of the former Delamerian orogen in western Victoria.

Discussion

Accretionary orogens are major sites of continental growth, reconstruction, and world-class mineralization. They have been active over much of Earth history, and the modern circum-Pacific orogenic belts and marginal basin–arc systems show typical examples of many of the variations in processes and stages of development. Ancient examples include: Phanerozoic orogens such as the Tasmanides, Altides, and early history of the Appalachian orogen; Proterozoic belts such as the Yavapai and Central Plains orogens of North America, and the late Palaeoproterozoic to Mesoproterozoic orogens of central and western Australia; and possibly even some Archaean gneiss–granite–greenstone belts. The variations in geodynamic processes within accretionary orogens is very wide, with some consisting primarily of recycled and reworked older

Fig. 10. (*Continued*) between *c.* 490 and 470 Ma with detritus derived from the Delamerian–Ross orogen. At *c.* 485 Ma calc-alkaline volcanism in the Macquarie Island Arc initiated as a result of subduction of the palaeo-Pacific (Glen *et al.* 2007). The Macquarie Arc developed more than 2000 km off the Gondwanan margin based on palinspastic reconstructions (Gray *et al.* 2006*a*) At *c.* 460 Ma subduction initiated along both sides of the Lachlan marginal basin (Soesoo *et al.* 1997; Foster & Gray 2000). Oceanic thrust systems were active in both the eastern and western parts of the basin from *c.* 455 to 439 Ma (Foster *et al.* 1999). Oblique convergence at *c.* 410 Ma resulted in strike-slip motion on the shear zones bounding the Wagga–Omeo Metamorphic Belt (WOMB), and thrusting along the southern margin (Foster *et al.* 1999). Widespread post-orogenic magmatism in the western part of the Western Lachlan orogen at *c.* 400 Ma was followed by final closure of the marginal basin (Melbourne Zone) at *c.* 390 Ma (Foster *et al.* 1999). The collision between the Western Lachlan accretionary-style thrust belt and the Central Lachlan accretionary complex and magmatic arc was accompanied by localized strong to intense north–south folding and regional, meridional crenulation cleavage development in the Central Lachlan orogen (Morand & Gray 1999). In the Eastern Lachlan orogen periods of shortening separated longer intervals of extension and extension-related volcanism above the outer subduction zone between *c.* 440 and 420 Ma, along with high-*T* metamorphism and syndeformational granitic magmatism (e.g. Cooma Complex; Fergusson & VandenBerg 1990; Zen 1995; Collins 2002*b*). Closure of Early Devonian extensional basins formed within the older Macquarie Arc occurred between 410 and 390 Ma (Cobar Basin: Glen *et al.* 1992) and again between *c.* 360 and 340 Ma (Hill End Basin: Foster *et al.* 1999). Post-orogenic magmatism in the eastern part of the Western Lachlan (central Victorian magmatic province) occurred at *c.* 370–360 Ma, and was followed by post-orogenic magmatism in the Eastern Lachlan orogen between 360 and 340 Ma.

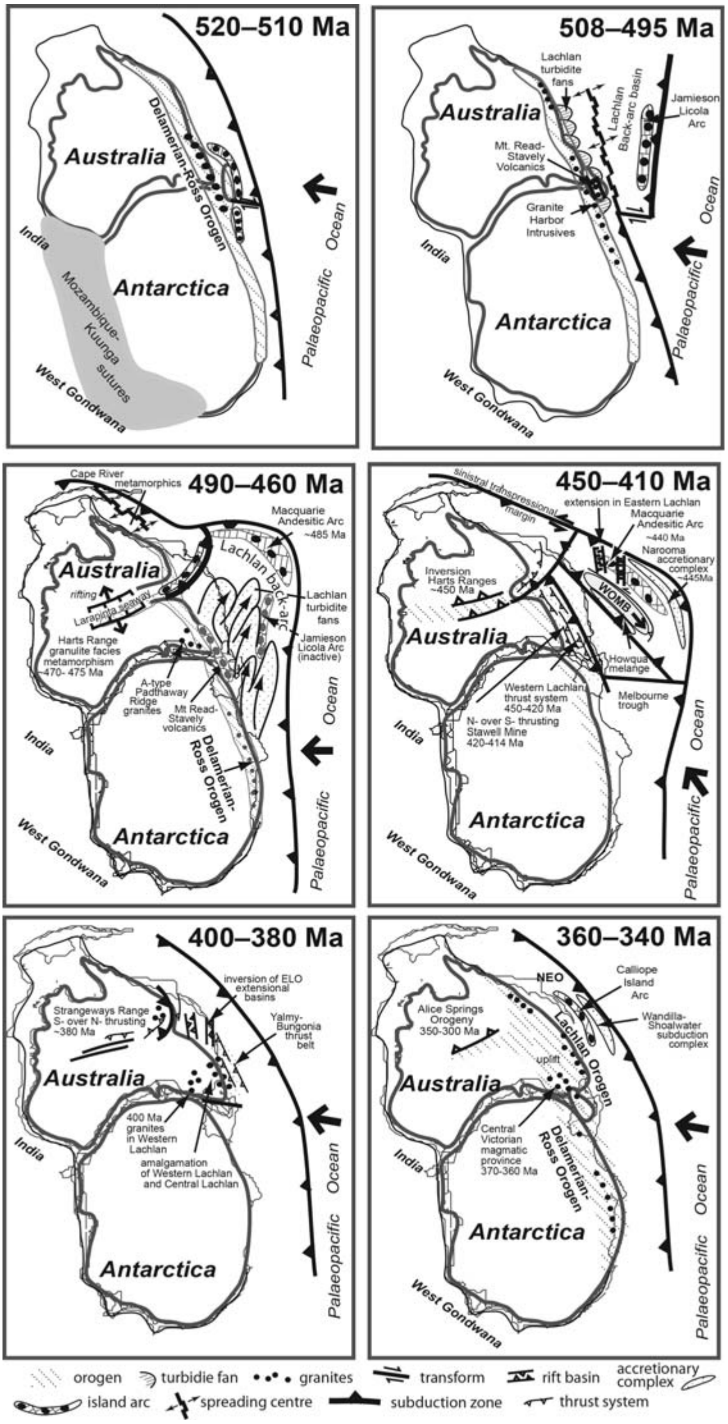

Fig. 10. Reconstruction of the tectonic evolution of the Lachlan orogen (after Foster & Gray 2000; Gray & Foster 2004). The Lachlan orogenic cycle started after collapse of the Delamerian–Ross orogen and subduction rollback opened the proto-Lachlan back-arc basin between *c*. 505 and 495 Ma (Foster *et al.* 2005). The Jamieson–Licola volcanic rocks were erupted during this time. Extensive turbidite fan deposition took place in the developing basin

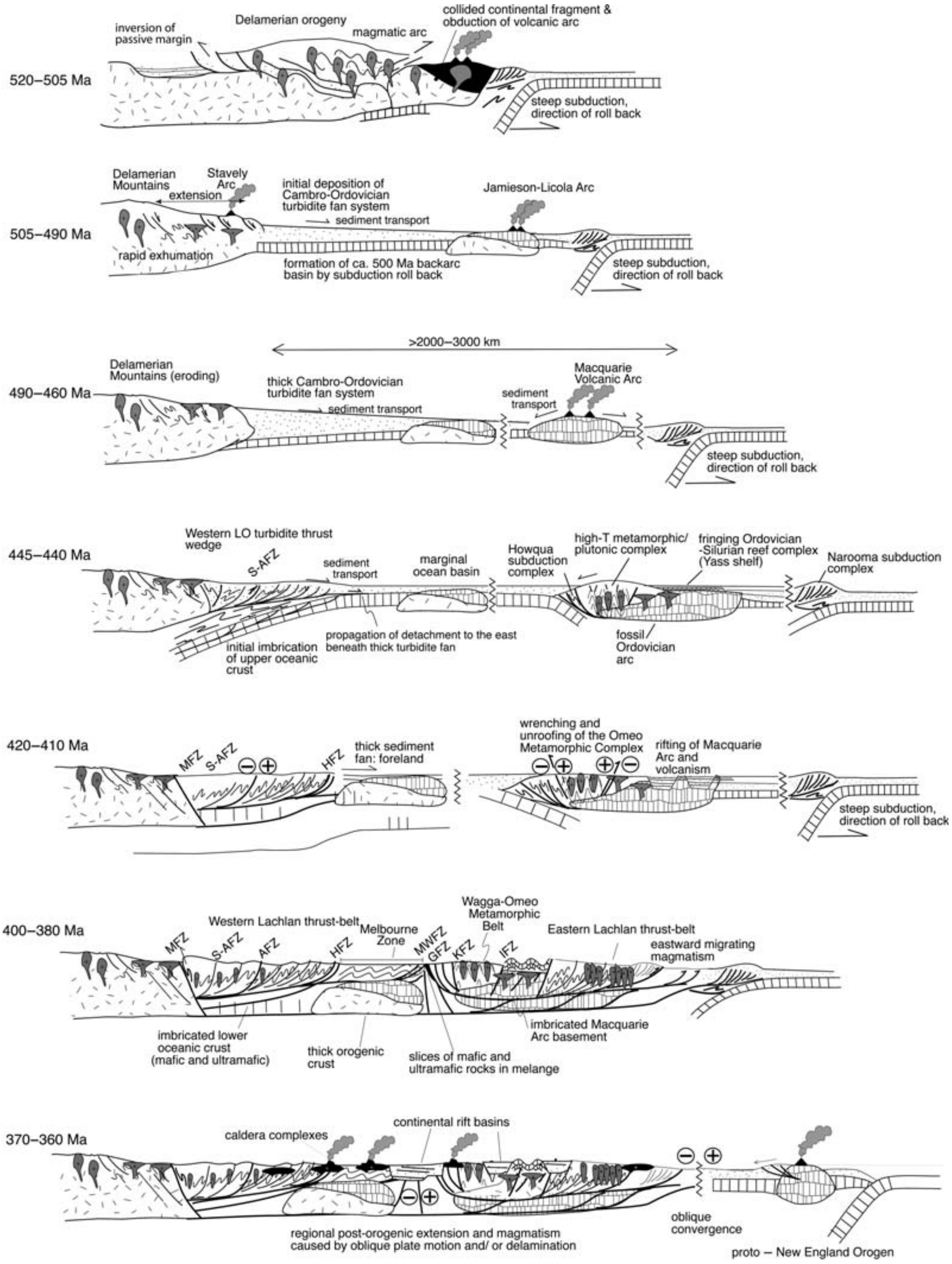

Fig. 11. Schematic cross-sections depicting the events and major tectonic elements that accreted to Australia to form the Lachlan orogen in Palaeozoic time (see Fig. 10 caption for a summary of key events). Abbreviations of major faults: AFZ, Avoca fault zone; GFZ, Governor fault zone; HFZ, Heathcote fault zone; KFZ, Kiewa fault zone; IFZ, Indi fault zone; MFZ, Moyston fault zone; MWFZ, Mt Wellington fault zone; SAFZ, Stawell-Ararat fault zone.

continental crust whereas others are constructed of almost entirely new continental crust composed of juvenile contributions from the mantle and structural thickening of oceanic materials. The most primitive accretionary orogens begin with oceanic elements combined with recycled continental detritus and many eventually become cratonized through thermal and chemical maturation processes. The Lachlan orogen is one of the type examples of a primitive accretionary orogen that began as a back-arc or marginal oceanic basin covered with a thick turbidite blanket, which was eventually closed by subduction processes (Fig. 10). Episodes of voluminous granitoid magmatism, driven by subduction and extension, eventually thermally and chemically matured the crust leading to cratonization. The Lachlan style of accretion fits many continental growth and recycling models going back through geological time to the Archaean (e.g. Condie 1982; Hamilton 1988; Coney 1992; Royden 1993; Sengor & Natal'in 1996; Foster & Gray 2000; Collins 2002*a*, *b*; Ingersoll *et al.* 2003; Percival *et al.* 2004). It is, however, critical to realize that long-term accretion in convergent settings does not typically occur simply through continuous subduction–accretion over hundreds of millions of years, but through a complex series of extensional events, some leading to marginal basins thousands of kilometres wide, interspersed with basin closure via subduction and shortening (e.g. Foster & Gray 2000; Collins 2002*b*; Gray & Foster 2004; Bickford & Hill 2007).

Most structural, geochemical, and geophysical data from the Lachlan and New England orogens suggest that they were constructed primarily on oceanic basement (reviewed by Gray & Foster 2004; Glen 2005). Whenever basement rocks to the Palaeozoic turbidite successions are exposed in the Lachlan orogen they consist of fault-bounded slices of Cambrian (and late Neoproterozoic?) mafic and ultramafic igneous rocks (Crawford 1988; Spaggiari *et al.* 2002*a*, *b*, 2003*a*–*c*, 2004; Crawford *et al.* 2003). There may be small intervening blocks of thin continental basement involved in accretion, but they are subordinate to the overall oceanic basin setting. Ordovician island arc volcanic rocks form the dominant basement of the Eastern Lachlan orogen, but these were probably constructed on oceanic crust (Glen *et al.* 2002, 2007). The extrapolation of aeromagnetic anomalies from Tasmania, granitoid generation models, and sedimentary facies models for the early Palaeozoic turbidites and black shales have led several workers to conclude that a larger block of Proterozoic continental crust underlies the central–southern Lachlan orogen (Scheibner & Veevers 2000; VandenBerg *et al.* 2000; Cayley *et al.* 2002; Clemens 2003). The new and previously published elemental and isotopic data from the Cambrian volcanic rocks that would have been erupted through this basement do not require the presence of Precambrian continental lower crust in this region, but do not totally rule it out. A structurally homogeneous and more felsic lower crust in this part of the Lachlan orogen could alternatively be the buried roots of a more extensive Cambrian Jamieson–Licola Arc assemblage.

In our interpretation, Western and Central Lachlan orogen accretion occurred by thickening and imbrication of extensive and thick submarine fans and the underlying marginal basic crust, in a Molucca Sea-style double-divergent subduction system (e.g. Hall *et al.* 1995). In this scenario subduction zones were initiated on the opposing sides of a former Cambrian–Ordovician back-arc basin (Fig. 11). Thrusting in the western Lachlan orogen resulted in marked shortening (*c.* 65–70%) by chevron folding, cleavage development and thrust imbrication of the turbidites. Subduction along the eastern side of the marginal basin, represented by the Howqua Accretionary Complex, produced large elongate composite batholiths within a shear zone-bounded, NW–SE-trending magmatic arc, as well as the associated high-temperature–low-pressure metamorphism. The Eastern Lachlan orogen is dominated by crust composed of the fragmented Ordovician Macquarie Arc, which was periodically rifted and then shortened above an outboard subduction zone. Mafic to felsic magmatism throughout the Lachlan orogen was generated by a variety of processes over time including subduction, extension, crustal thickening, and post-orogenic extension (Figs 10 and 11). The primitive nature of the initial tectonic elements of the Lachlan orogen also explains why this belt hosts a diverse range of rich mineral deposit types (e.g. Bierlein *et al.* 2002, 2006).

This research was funded by grants from the Australian Research Council and National Science Foundation. C. Fergusson and R. Korsch are thanked for providing helpful reviews of the manuscript.

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