Fragmented Tasmania: the transition from Rodinia to Gondwana

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The origin of the microcontinent VanDieland extends back to the late Paleoproterozoic, where it was positioned between East Antarctica and southwestern Laurentia, within the supercontinent Nuna and Rodinia. Paleo- to Mesoproterozoic events recorded in VanDieland have greater affinities with southwest Laurentia and East Antarctica, suggesting southern VanDieland was part of the Grenville Front, and the central Tasmanian part was adjacent to the Miller Range in the central Transantarctic Mountains. Late in the Neoproterozoic Rodinia break-up, VanDieland separated from East Antarctica and southwestern Laurentia, and moved north along the Terra Australis margin until its southern part was positioned next to the easternmost Robertson Bay Terrane of north Victoria Land. VanDieland comprises up to seven different crustal megaboudins or microcontinental ribbon terranes that likely had amalgamated by the end of the Cambrian; these ribbon terranes are bounded by major faults and suture zones. Some boundaries, such as the Arthur Metamorphic Complex, are well known. However, other boundaries, like the eastern edge of the Tyennan Zone, and the boundary between King Island and northwestern Tasmania, are more cryptic, as they are covered by younger geology or are under water. The boundaries are commonly defined by sedimentary and mafic volcanic infill that has been trapped between the crustal fragments. These rocks have previously been interpreted as allochthonous terranes but are more likely to represent inverted sections of attenuated transitional crust and back arc basin fill that formed along the eastern margin of the Gondwana plate during the Cambrian. This interpretation also provides an explanation for the previous tectonic analysis that suggests that Tasmania’s mafic- ultramafic complexes were obducted westward onto older sequences and were subsequently transported southwards as other ribbons collided along the northeastern and western edges of the growing microcontinent, which existed in the overriding plate of a west-dipping subduction zone at the convergent margin between Gondwana and the proto-Pacific plate.

KEYWORDS: Western Tasmania, Rodinia, Gondwana, supercontinent, VanDieland, Tasmanides.

INTRODUCTION

The period from 600 Ma to 500 Ma includes some of the most fundamental changes in Earth history. Not only did animals develop hard skeletons and the ‘snowball earth’ time finish, but the Rodinian supercontinent had completely dispersed and the Gondwana supercontinent began to amalgamate. Facing the exterior ocean of Gondwana was the Terra Australis Orogen (Cawood 2005), which records the accretion of dismembered fragments and oceanic material at the edge of Gondwana. This paper examines the geological evolution and crustal architecture of western Tasmania. The region contains excellent evidence of some of the last breakup events of the remnants of Rodinia, at about 580 Ma (Calver et al. 2004; Meffre et al. 2004). Thus, western Tasmania provides an opportunity to examine a fundamental transition in Earth history, from almost 300 Ma of continent destruction to 300 Ma of continent building. Furthermore, the Cambrian orogenesis in Tasmania apparently took place isolated from the rest of Gondwana, and terranes of western Tasmanian were not truly accreted onto the Gondwanan margin until the Middle Devonian (Cayley et al. 2002, 2011; Moresi et al. 2014). Hence, the region provides opportunity to understand tectonic processes associated with Rodinia break-up and subsequent accretion on the Gondwanan margin during the Cambrian to Devonian stages of the Terra Australis Orogen.

Cayley (2011) coined the term ‘VanDieland’ for a Proterozoic micro-continent that included western Tasmania, the Selwyn Block (central Lachlan Fold Belt), the South Tasman Rise and the East Tasman Plateau. VanDieland extends over 1500 km from north to south (Figure 1) and has been considered to represent a single, coherent entity (Berry et al. 2007; Cayley 2011; Berry & Bull 2012). Western Tasmania (Figures 1, 2) contains the most complete and accessible geological record of VanDieland. This region contains a wide range of pre-Ordovician rocks, including deep-water turbidites, non-marine and shallow marine clastic deposits, carbonate

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Figure 1 (A) Southern Victorian and Tasmanian geology, showing the near-surface boundaries of the Selwyn Block and Tasmanian Elements (from Seymour & Calver 1998). Elements are labelled as: AJ, Adamsfield–Jubilee; D, Dundas; ET, Eastern Tasmania; KI, King Island; S, Sheffield; RC, Rocky Cape; TY, Tyennan. Inset shows the location within Australia. (B) Tasmanian geology as in (A) with the total magnetic intensity as an intensity layer. Suggested zones are BU Burnie Zone; ET Eastern Tasmania; KI, King Island; SB, Sorell–Badger Head; P, Pedder (including the Adamsfield–Jubilee Element; see sections on the Pedder Zone and the origin of the mafic–ultramafic complexes); RC, Rocky Cape; T, Tyennan Zone. B, Braddon River Fault; G, the Tamar Graben; H, the Mt Hobhouse Fault. (C) Upward continued total magnetic intensity data. Lines and lettering as for (B). (D) Schematic section across Tasmania.
The oldest rocks were deformed and metamorphosed in the Mesoproterozoic (Chmielowski 2009). Extensive areas of VanDieland contain Neoproterozoic to Cambrian rift assemblages of deep-water metasedimentary rocks, granitic plutons, within-plate basalts or ocean floor basalts, Cambrian boninite-associated mafic–ultramafic complexes, eclogite, blueschist, greenschist, and amphibolite facies metamorphic rocks (Berry & Crawford 1988; Crawford & Berry 1992; Black et al. 1997; Corbett et al. 2014). Low-temperature–high-pressure metamorphic rocks formed during the middle to late Cambrian Tyennan Orogeny (Meffre et al. 2000; Chmielowski & Berry 2012). Rocks with oceanic affinities have been interpreted as allochthonous sheets emplaced at one or more collision zones associated with the Tyennan Orogeny (e.g. Crawford & Berry 1992; Turner et al. 1998; Meffre et al. 2000; Stacey & Berry 2004; Berry et al. 2005; Berry & Bull 2012; Chmielowski & Berry 2012; Seymour et al. 2013). The presence of these obducted mafic–ultramafic rocks suggests that discrete continental crustal blocks may have been separated by oceanic or transitional crust (Moore et al. 2013).

The Tyennan Orogeny is one example of the earliest phases of Gondwanan orogenesis (Crawford & Berry 1992; Everard et al. 2007; Berry & Bull 2012; Seymour et al. 2013) and appears to be part of a much larger orogenic system that includes the Ross Orogeny in Antarctica and the Delamerian Orogeny on mainland Australia (Boger & Miller 2004; Cawood 2005; Champion et al. 2009; Boger 2011). By the Ordovician, the extensive limestone sheets of the modern northwest Pacific (Cawood et al. 2009). Our proposed model offers tectonic solutions for longstanding paradoxes such as the apparent change in Cambrian subduction polarity along the Gondwana margin (cf. Crawford & Berry 1992; Miller et al. 2005; Foden et al. 2006), as well as an opportunity to assess and test different tectonic models, including those involving rapid changes of subduction polarity (Crawford & Berry 1992; Crawford et al. 2003b), oblique subduction on an outboard fragment (Cawood 2005), and complex tectonic scenarios where contemporaneous east- and west-dipping subduction zones were present in the early Cambrian and VanDieland was sucked obliquely into Gondwana by the west-dipping subduction zone in the Late Cambrian before rotating and moving away from Gondwana during the Early Ordovician (Cayley 2011). Finally, we place VanDieland in the context of Precambrian continental reconstructions.

### REGIONAL GEOLOGY

Tasmania is typically separated into two regions. Western Tasmania comprises Mesoproterozoic and Neoproterozoic cratonic crust, whereas eastern Tasmania has oceanic affinities and was cratonised in the Middle Devonian Tabberabberan Orogeny (Figures 1, 2) (e.g. Reed et al. 2002; Black et al. 2004; Berry & Bull 2012; Seymour et al. 2013). The boundary between eastern and western Tasmania is believed to be present beneath the Tamar Graben (Figure 1b) (Seymour et al. 2013), just west of which is seen a Lower Devonian sequence that has affinities with both eastern and western Tasmanian rocks (Rickards et al. 2002). Further south, the boundary is covered by Permian and younger rocks. Offshore seismic reflection data from eastern Tasmania indicated the presence of Proterozoic crust at depth along the southern half of the Tasmanian east coast (Drummond et al. 2000). In contrast, Seymour et al. (2013) suggested that eastern Tasmania is not floored by Proterozoic crust and placed the boundary further west. Eastern Tasmania has been widely considered as a single block or ‘element’ (Burrett & Martin 1989; Seymour & Calver 1998; Corbett et al. 2014).

Western Tasmania records a more complex history including Neoproterozoic rifting, which involved break-up and eastward drift of the micro-continent from East Antarctica (Berry et al. 2008), followed by middle to late Cambrian crustal shortening and subsequent...
extensional tectonism, the Tyennan Orogeny (Berry & Bull 2012). Late Ordovician to Devonian reworking associated with the accretion of VanDieland into the eastern Gondwanan margin (Cayley et al. 2011) affected both eastern and western Tasmania as eastern Tasmania was either partly (Patison et al. 2001) or completely (Powell & Baillie 1992; Reed et al. 2002) transported westward during the Tabberabberan Orogeny. During the Jurassic, a Large Igneous Province resulted in voluminous emplacement of dolerite (Hergt & Brauns 2001). Cretaceous and Paleogene extension and crustal thinning led to the break-up of Australian and Antarctica (Norvick & Smith 2001; Veevers 2012), which was followed by Cenozoic basaltic volcanism and sedimentation (Stacey & Berry 2004; Data Management Group 2011; Corbett et al. 2014).

Seymour & Calver (1995) divided western Tasmania into six ‘elements’: King Island, Rocky Cape, Dundas, Sheffield, Tyennan and Adamsfield–Jubilee (Figure 1a). Some boundaries between the elements are well established. For example, the Arthur Metamorphic Complex is a high-strain zone containing blueschists and rift-related rocks that separate the Rocky Cape and Dundas Elements (Turner & Bottrill 2001; Holm & Berry 2002). Other boundaries are covered. The boundary separating eastern Tasmania from elements to the west is beneath younger sedimentary successions, Jurassic dolerite and Cenozoic basalt. The boundary between King Island and the Rocky Cape Element is covered by water; and Cretaceous and Cenozoic sedimentary rocks. The boundaries have also been modified by later events, notably the
Devonian Tabberabberan Orogeny (Holm & Berry 2002; Reed et al. 2002). The multiple deformations resulted in significant metamorphic grade variations and styles across the boundaries, which range from greenschist to amphibolite and blueschist facies (Meffre et al. 2000; Chmielowski & Berry 2012). Age dating criteria are also problematic as most record ages that cluster at ca 510 Ma (e.g. Berry et al. 2007, figure 10).

The following section first outlines each of the zones that comprise western Tasmania and the South Tasman Rise. Figure 2 shows the pre-Silurian stratigraphic columns for each zone. Some zones replicate the “Elements” of Seymour & Calver (1995, 1998), but we have chosen to vary or subdivide others. In order to avoid confusion, we have called our divisions “Zones” and only used the term “Element” where it refers to areas defined by them.

**King Island Zone**

King Island largely comprises Mesoproterozoic metasedimentary rocks with detrital zircon populations older than 1350 ± 90 Ma (Black et al. 2004). In the west, the rocks underwent regional metamorphism to greenschist or amphibolite facies at 1287 ± 18 Ma (U/Th/Pb on monazite; Berry et al. 2005). On the west and north coasts, later metamorphism, deformation and granite intrusion at 760 ± 12 Ma is coeval with ca 750–780 Ma granites (Black et al. 1997; Turner et al. 1998; Calver et al. 2013b). On the east coast, metasedimentary successions are unconformably overlain by Marinoan glaciomarine deposits with detrital zircon dates of ca 636 Ma (chemically abraded TIMS 238U/206Pb zircon; Calver et al. (2013a). This sequence is successively overlain by a cap carbonate, laminated black shale and the mafic volcanic (rift tholeiites) and intrusive rocks that form the eastern edge of the zone. This package of rocks is magnetic and relatively dense and is evident in regional geophysical data as a 35 km-wide zone (Figures 1, 3–5) that extends between outcrops on the east coast of King Island to Phillip Island on the southern Victorian coast (Henry & Birch 1992; Moore et al. 2013). The magnetic package is truncated by the Braddon River Fault, approximately 40 km to the south of King Island. The package dips
Figure 3 Tilt-derivative filtered magnetic image of Tasmania and adjacent areas, together with the zone boundaries (black), later faults (yellow) and letter symbols from Figure 1b.

Figure 4 Gravity image of Tasmania and adjacent areas. (a) Bouguer onshore, free air offshore, (b) Isostatic onshore, free air offshore, together with the zone boundaries (black), later faults (yellow) and letter symbols from Figure 1b.
moderately to steeply to the east between 45° and 70° (Meffre et al. 2004). Meffre et al. (2004) suggested the sequence was fault repeated, and proposed a pre-repetition width of several tens to hundreds of kilometres of rift tholeiites. Based on stratigraphic and geochemical considerations, a volcanic passive margin sequence setting was interpreted. Sm/Nd analyses suggested the tholeiites were emplaced at ca 580 Ma (Meffre et al. 2004). A TDM model age of 1700 Ma was obtained from the most evolved samples (ƐNd/ε0 = 3.1 at 579 Ma; Meffre et al. 2004). Mafic dykes in the sequence yielded a U/Pb SHRIMP date of 575 ± 3 Ma (Calver et al. 2004).

The Braddon River Fault (Figures 1, 3–6) is prominent in regional magnetic images of offshore western Tasmania. It is exposed onshore at the Braddon River headwaters, where late movement has displaced early Paleozoic rocks (Baillie & Corbett 1985). Further south, it is defined by a 2 km-wide mylonite at the eastern edge of a Neoproterozoic mafic volcanic rock package (Brown 2011). The Braddon River Fault is imaged in seismic reflection data as steeply west-dipping and has a significant change in seismic character across it (Figure 7). South of King Island, it appears to have a displacement of approximately 70 km. However, further south much of this movement appears to be taken up in internal faulting within the Burnie Zone (Figure 8).

Rocky Cape Zone

The Rocky Cape Zone occurs in the northwest corner of Tasmania and the northern Sorell Peninsula (Figures 1–4, 6, 8, 9). The oldest rocks are the Mesoproterozoic Rocky Cape Group, which comprises alternating sequences of marginal marine quartzite and shelf siltstone (Figure 9; Table 1; Everard et al. 2007; Halpin et al. 2014). The presence of the fossil Horodyskia williamsii suggests sedimentation at the basal part of the package occurred between 1400 and 1100 Ma (Calver et al. 2010). Authigenic monazite from quartzites yield dates that cluster in three populations, ca 1360–1290 Ma, ca 1280–1240 Ma and ca 1090 Ma (Halpin et al. 2014). Cambrian and Devonian granites that intrude the Rocky Cape Group contain slightly different zircon inheritance patterns, notably an excess population of 1650–1600 Ma ages that are poorly represented in the Rocky Cape Group (Black et al. 2010). These are likely to be sourced from the cratonic basement on which the Rocky Cape Group was deposited (Black et al. 2010).

Within the Smithton Basin, the Togari Group unconformably overlies the Rocky Cape Group (Everard et al. 2007). It includes conglomerate, dolomite, chert, diamicrite, volcaniclastic rocks, siliceous metasedimentary rocks and tholeiitic basalt, which were deposited in an
extensional setting (Everard et al. 2007). $\delta^{13}$C values and $^{87}$Sr/$^{86}$Sr ratios (Calver 1998) from the Togari Group and correlations with similar succession of the Adelaide Fold Belt suggest the lower Togari Group may be mid-Cryogenian (Everard et al. 2007). A rhyodacite in a rift tholeiite package in the upper Togari Group yielded an U/Pb zircon SHRIMP age of 582 ± 4 Ma (Calver et al. 2004), while the uppermost siltstone unit contains early to middle Cambrian fossils (Everard et al. 2007). An upper Cambrian marginal marine to terrestrial clastic sequence overlies the Togari Group. This package contains serpentinite-rich detritus from mafic–ultramafic rocks, similar to those that were obducted onto the adjacent Burnie Zone during the Cambrian Tyennan Orogeny (Everard et al. 2007).

The Rocky Cape and Togari groups are also present on the Sorell Peninsula (Figure 8; Corbett 2003). We suggest that the Sorell Peninsula rocks have been displaced from the rest of the Rocky Cape Zone by approximately 50 km of sinistral movement on the Braddon River Fault (Figure 11).

**Burnie Zone**

The oldest rocks of the Burnie Zone are the clastic metaturbidites of the Burnie (Oonah) Formation (Figures 10–13). Similar detrital zircon populations to the shallow water Rocky Cape Group can be interpreted as the two being coeval (Black et al. 2004). More likely, the Burnie Formation represents a younger package derived from the Rocky Cape Group or similarly sourced rocks. The minimum age for the Burnie Formation is ca 710 Ma based on a K-Ar biotite date in an alkaline dolerite that was intruded into wet sediments (Figure 13b; McDougall & Leggo 1965, recalculated in Black et al. 2004), suggesting the Burnie Formation correlates with the basal successions of the Togari Group in the Rocky Cape Zone (Calver 1998).

Further south, the Burnie Formation is overlain by a shallow water rift sequence of the Success Creek Group, and the deep-water turbidites and tholeiitic basalt, chert and mudstone of the Crimson Creek Formation (Brown 1986). The Success Creek Group has been correlated with ca 700 Ma succession in the Smithton Basin based on similarities in the stromatolites present (Brown 1986). The Crimson Creek Formation is considered to have been deposited on a rifting margin (Brown 1986), with the basalts interpreted as seaward-dipping reflectors (Direen & Crawford 2003).

Boninitic mafic–ultramafic complexes within the Burnie Zone include dunite–harzburgite, layered pyroxenite–dunite or layered peridotite–pyroxenite–gabbro
cumulates that are interpreted to have formed in a forearc settings (Crawford & Berry 1992; Stern et al. 2012) or west-directed obducted slices from supra-subduction zones during the early phase of the Tyennan Orogeny (Stacey & Berry 2004). These rocks represent a small component of the Burnie Zone (Berry & Crawford 1988). The ages of these boninitic complexes are mostly poorly constrained, although Mortensen et al. (in press) gave a U/Pb zircon ca 516 Ma date for an associated gabbro in the Burnie Zone. The packages mark a gross change in tectonic environment from crustal extension to crustal shortening.

During the middle Cambrian, in the second phase of the Tyennan Orogeny, the marine succession of the Mount Read Volcanics was deposited in north-south graben associated with east-west extension (Berry & Bull 2012). The Mount Read Volcanics vary in composition from basalt to rhyolite and include intrusions, lavas and volcaniclastic metasedimentary rocks, shelf limestone and mudstone (Burrett & Martin 1989; Corbett et al. 2014) deposited between ca 507 and ca 495 Ma (Seymour et al. 2013; Mortensen et al., in press). Berry et al. (2008) gave a $T_{\text{DM}}$ of approximately 1670 Ma from two felsic porphyries, suggesting a late Paleoproterozoic mantle extraction age.

Figure 7 Seismic line, AGSO 148/9, upper panel shows the interpretation, lower panel uninterpreted line, inset gives the line location. The Braddon River Fault is sub-parallel to the seismic line causing significant off-line effects. Shot points are approximately 50 m apart.
In the upper Cambrian, the Owen Conglomerate and associated rocks were deposited in half graben (Noll & Hall 2005). By the Middle Ordovician, sedimentation had evolved from coarse clastics to fine clastics and then to the micritic dolomitic limestone of the Gordon Group (Seymour et al. 2013).

Pedder Zone

The Pedder Zone and the Tyennan Zone form the Tyennan Element of Seymour & Calver (1998) (Figures 1, 2, 14–16). The Pedder Zone is differentiated from the Tyennan Zone by the presence of the eclogite facies rocks of the Franklin Metamorphic Complex (Figure 15a). Metamorphic analysis of these eclogites yielded pressures between 1400 and 1960 MPa (depths of ~40–60 km) and temperatures of ~550–650 °C (Chmielowski & Berry 2012), and so are likely to have formed at a boundary between cratonic blocks rather than within the interior of a single block.

Calver et al. (2006) interpreted the oldest rocks in the Pedder Zone as clast-bearing proximal turbidites. A mylonite within this package yielded a typical 1900 to 1400 Ma detrital zircon population, but some had metamorphic overgrowths that yielded a date of 1220 ± 36 Ma (Chmielowski 2009), suggesting the metaturbidites were either metamorphosed at 1220 Ma, or derived from rocks that had been metamorphosed then. The shallow marine to subaerial succession of the Clark Group overlies these metaturbidites, and the similarities in lithologies and detrital zircon populations suggest correlation with the Rocky Cape Group (Black et al. 2004; Calver et al. 2006). Several samples contained a metamorphic monazite population of 1367 ± 7 Ma (Chmielowski 2009). Metaquartzite, dolomite and diamictite are in fault contact with, and are interpreted to overlie the Clark Group. These rocks are correlated with the Togari Group (Calver et al. 2006).

The siliceous Wings Sandstone, in the eastern part of the Pedder Zone, is characterised by an unusual detrital zircon population with dominant populations between ca 1400 and ca 900 Ma, with the youngest population 914 ± 44 Ma (Black et al. 2004). The Ediacaran to Cambrian lithic volcanic metasediments, chert and minor basaltic tuff of the Ragged Basin, and slices of ultramafic rocks occur adjacent to the Wings Sandstone (Calver et al. 2006). There are no direct age controls on these east-dipping ultramafic slices; however, Crawford & Berry (1992) suggested they were obducted at the same time as other Tasmanian mafic–ultramafic–boninite package at the western edge of a Cambrian felsic volcanic sequence. We interpret this fault system as an east-dipping, leading imbricate fan (Boyer & Elliott 1982). In the magnetic and seismic images, these major boundary systems can be traced south from the Braddon River Fault until they are truncated by the southern extension of the eastern boundary of the King Island Zone (Figures 3, 10). Although the mafic–ultramafic–boninite package has been subsequently disrupted by the Devonian Tabberabberan deformation, we consider that this western boundary of the Burnie Zone on the Sorell Peninsula is more or less in place. In the north, the Braddon River Fault has a sinistral displacement of up to 70 km. However, the eastern boundary of the Burnie Zone is only displaced by approximately 5 km, and we suggest that most of the missing movement has been taken up along the sheared eastern boundary of a Neoproterozoic mafic volcanic package within the Burnie Zone. These volcanic rocks can be seen as both magnetic and gravity highs.
The oldest Tyennan ages of metamorphic monazites in the Franklin Metamorphic Complex are 529 ± 10 Ma from a muscovite–quartz–garnet–plagioclase schist, and the youngest 505 ± 7 Ma from a quartz–muscovite–plagioclase–garnet–biotite schist; most zircon and monazite ages cluster at approximately 510 Ma (Chmielowski & Berry 2012; Fergusson et al. 2013). On the southwest coast near Nye Bay (Figure 12b), the metamorphism is slightly younger, with monazite ages clustering at 505 ± 2 Ma (Chmielowski & Berry 2012).

Tyennan Zone

The oldest mapped rocks in the Tyennan Zone are marginal marine metaquartzite and garnet schist, with a maximum deposition age of 1574 ± 59 Ma based on the youngest detrital zircon population (Black et al. 2004). The basement to this package is unknown (Calver et al. 2006; Black et al. 2010). Metamorphic conditions for these rocks reached 1960 MPa and 545 °C at 508 ± 9 Ma (U/Th/Pb monazite age; Chmielowski 2009) in a region near our interpreted boundary with the Pedder Zone. Drilling in central Tasmania intersected dolomite that was correlated with dolomite in the Togari Group of the Rocky Cape Zone (Figure 14; Reid et al. 2003).

The metaquartzites are overlain by metaturbidites that have previously been interpreted as Burnie Formation or its equivalents (Vicary et al. 2008). Along the north coast, the poly-deformed Goat Island Conglomerate, within the Burnie Formation, contains a large proportion of quartzite cobbles to boulders (Figure 13e; Berry & Gray 2001), suggesting, at least locally,

Table 1 Rocky Cape Group age constraints.

| Youngest sample | Jacob Quartzite | Detention Subgroup | Data sources |
|-----------------|-----------------|---------------------|-------------|
| Detrital zircon populations (Ma) | 1009 ± 43, 1250, 1440, 1850, 2650 | 1433 ± 14, 1690–1740, 1780 | Black et al. 1997, 2004 |
| Authigenic monazite (Ma) | | 1085 ± 9 | Halpin et al. 2014 |

| Oldest sample | Pedder River Siltstone |
|---------------|-----------------------|
| Detrital zircon populations | 1442 ± 28, 1630, 1720, 1790, 1850 |
| Authigenic monazite (Ma) | 1315 ± 23, 1350 |
| | | | Halpin et al. 2014 |

Figure 9 Rocky Cape Group (Detention Metaquartzite) with herringbone crossbeds just above the scale. Location 40°51’41“S, 145°30’44”E. Scale is 16 cm long.
significant paleo-topography and erosion from the quartzites now exposed to the south. The timing of deposition of this package is uncertain but pre-dates metamorphism and deformation at ca 510 Ma (Chmielowski & Berry 2012). Also present on the north coast of Tasmania are tectonic mafic/ultramafic complex rocks, and MORB basalt (Motton Spilite) that overlies Cambrian chert (Seymour & Vicary 2010; Everard & Calver 2014).

The Tyennan Zone was extensively deformed and metamorphosed during the Tyennan Orogeny, and formed the Ulverstone/Forth Metamorphic Complex. The Forth Metamorphic Complex was metamorphosed to peak temperatures of 700°C and pressures of 1690 MPa (Meffre et al. 2000; Chmielowski & Berry 2012). Metamorphism has been dated at 512 ± 5 Ma (SHRIMP; Black et al. 1997, recalculated by Foster et al. 2005) and 508 ± 2 Ma (40Ar/39Ar; Foster et al. 2005). The metamorphic grade increases in the footwalls of west-dipping faults and Berry & Gray (2001) suggested the entire sequence was inverted and allochthonous. We are not persuaded by this interpretation. Their mapping shows the metamorphic rocks in faulted contact with weakly metamorphosed packages, implying that the metamorphism took place before the faulting. If there were inversion, the sequence most likely came from the west. The Arthur Metamorphic Complex, at the western edge of the Burnie Zone, is characterised by temperatures and pressures of 700°C and 1690 MPa.
pressures of 350°C and 700 MPa (Turner & Bottrill 2001), which are significantly lower than those of the Forth Metamorphic Complex, suggesting they are not equivalents; other alternatives seem equally unlikely.

The Fossey Mountain Trough lies between the northem, coastal outcrops and the main Tyennan Zone in central Tasmania. This rift is filled with Mount Read Volcanics equivalents that are overlain by Owen Conglomerate and equivalent sequences. These are overlain by the Ordovician Gordon Group (Data Management Group 2011). North of the rift, there appears to be a broad facies change in the basement rocks, with the original sediments apparently less quartzite-rich. However, metaquartzite schist and dolomite are present in some of the north-easternmost exposures (McClenaghan & Vicary 2005), and these can be correlated with rocks to the south of the rift. The isostatic gravity data (Figure 4) suggest that meridional pre-rift structures can be traced north under the rift. Elsewhere, drilling near Hobart intersected high-Al basalt that has similar geochemistry and petrology to basalt in the Mount Read Volcanics (Crawford & Berry 1992).

**Sorell—Badger Head Zone**

Metaturbidites at Badger Head, in the eastern part of this zone (Figures 14, 15b, 17), contain a youngest detrital zircon population of 1242 ± 29 Ma (Black et al. 2004),
implying syn- or post-Grenville Orogeny deposition. The rocks have been correlated with the Burnie Formation turbidites of the Burnie and Tyennan zones (Gee & Legge 1979). They are metamorphosed to lower greenschist facies, and locally contain retrogressed garnet (Reed et al. 2002).

To the west of these metaturbidites are mudstone, dolostone, volcaniclastic sandstone, conglomerate, chert, dolerite, rift tholeiite and rare rhyolite. These rocks are also associated with bedding-parallel broken formation (Calver & Reed 2001). Microfossil-bearing chert indicates deposition during the Cryogenian, which is supported by late Neoproterozoic $\delta^{13}$C values from these rocks (Calver & Reed 2001). The sequence may correlate with the Crimson Creek Formation of the Burnie Zone (Reed et al. 2002).

The Andersons Creek Ultramafic Complex lies immediately east of the metaturbidites at Badger Head (Figure 15b). Magnetic and gravity modelling suggested the complex lay in the hangingwalls of east and west-vergent thrusts. The west-dipping thrust shallows beneath the Badger Head rocks, whereas the east-dipping thrust extends beneath Cambrian to Devonian platform sequences (Zengerer 1999). Highly deformed granite in the apex of the ultramafic complex yielded U–Pb SHRIMP dates of $638 \pm 5$ Ma and $661 \pm 8$ Ma (Black 2007, D. H. Moore et al. 2003).
unpublished data, OZCHRON database). The intrusion was metamorphosed to amphibolite facies at ca. 510 to 520 Ma (U/Th/Pb on monazite; Berry et al. 2007).

Reed et al. (2002) proposed a complex tectonic model involving two intersecting subduction zones during the Tyennan Orogeny. The model assumes that the rocks of the Sorell–Badger Head Zone were allochthonous and not connected to western Tasmania prior to the Tyennan Orogeny. In this model, Reed et al. (2002) suggested the change between eastern and western Tasmania is a

Figure 13 (a) Simple anticline in the Burnie Formation, outlined by a dark shale bed. Location 41°4’9″S, 145°56’22″E, on the eastern edge of Burnie. (b) Lighter-coloured Burnie Formation intruded by dark dolerite; the Burnie Formation has been bleached by the interaction with water heated by the dolerite intrusion. Location 41°2’35″S, 145°53′E, east of Cooee. K/Ar dating on biotite from similar mafic rocks gave an age of 711 ± 16 Ma. (c) Complexly deformed Burnie Formation just west of the Tyennan–Burnie boundary. White scale is 16 cm long. Location 41°6’16″S, 146°3’51″E, west of Penguin. (d) Multiple episodes of quartz veining in the Penguin Fault (at the Tyennan–Burnie boundary) some of which show a sinistral movement sense. Brecciation and a network of quartz veins is present over approximately 270 m across the dominant strike of veining. Location 41°6’19″S, 146°3’56″E, approximately 150 m southeast of (c). (e) Folded conglomerate and sandstone east of the Tyennan–Burnie boundary; view is to the south. Location 41°8’12″S, 146°8’12″E, south of Goat Island, approximately 3 km west of Ulverstone.
simple facies change from shallow-water succession in the west to deep-water succession in the east. This model does not account for the presence of the 658 Ma granite in the footwalls of both thrusts; nor is it easily reconciled with the geology of the other tectonic zones throughout Tasmania and the regional west-dipping subduction system outboard of Tasmania (Cawood 2005; Squire & Wilson 2005; Stump et al. 2006). We accept the interpretation that the Badger Head metaturbidite package is allochthonous (Reed et al. 2002) and has been thrust eastwards above the ultramafic rocks. However, we prefer the interpretations by Powell & Baillie (1992) or Patison et al. (2001) that infer the presence of concealed Precambrian cratonic crust east of the Andersons Creek Ultramafic Complex as a basement to the Cambrian and Ordovician shelf sequences there, and consider that the eastern boundary of the cratonic crust lies under the Tamar Graben (Rawlinson et al. 2010; Young et al. 2011).

Glomar Zone

Royer & Rollet (1997) and Berry et al. (1997) first outlined the boundaries, nature and positions of the thinned continental crust that lies to the south of Tasmania. Based on the submarine topography, Royer & Rollet (1997) subdivided the region into three areas, the west South Tasman Rise, the east South Tasman Rise and the East Tasman Plateau (Figure 18). Reconstructions by Exxon et al. (1997a), Royer & Rollet (1997) and Norvick & Smith (2001) indicated that the East Tasman Plateau was rifted from the east South Tasman Rise in the Late Cretaceous.

In order to avoid confusion with their terminology, we have grouped a slightly redefined east South Tasman Rise and the East Tasman Plateau into the Glomar Zone, after the ship that recovered the first pre-Mesozoic rocks in the region. Exxon et al. (1995) briefly described the results from four cruises in the region, three of which had recovered pre-Mesozoic samples from 42 dredge sites or drill holes. Subsequent age dating on 14 samples has allowed initial conclusions to be drawn about the South Tasman Rise.

The east South Tasman Rise is topographically smoother, and the bathymetry is ~1000 m. Seismic interpretations indicate the presence of up to 2 s TWT (~3 km) thickness of Cretaceous and Cenozoic sedimentary rocks (Exxon et al. 1997b). Basement is interpreted to have been attenuated during Gondwana break-up; however, the present-day expression is consistent with separation during the Tasman rift event. The free air gravity response has a strong linear pattern that trends at about 120°, although along its western edge, the gravity anomalies trend parallel to the boundary with the west South Tasman Rise.

**Figure 14** Central Tasmania, showing the Pedder, Tyennan and Sorell – Badger Head zones. (a) Geology with a greyscale background of the total magnetic intensity; colours as for Figure 1. Inset gives the location of the images. (b) Tilt filtered total magnetic intensity. (c) Isostatic gravity onshore, free air gravity offshore. H marks the Mt Hobhouse Fault, W the Mt Wedge Fault, the western boundary of the Wings Subzone. The boundary between eastern and western Tasmania is constrained by drill holes at G that intersected andesite similar to that in the Mount Read Volcanics (Crawford & Berry 1992), at M that intersected sheared Mathinna Group (Clarke & Farmer 1983) and at X that intersected probable Proterozoic dolomite (Reid et al. 2003). The western boundary of eastern Tasmania is also indicated by the rise of approximately 60 μm s⁻² in the isostatic gravity response in western Tasmania. The boundary is also visible in the offshore seismic data as a feature that dips east at approximately 30° from 2 to 8 s TWT (approximately 6–24 km depth, Figure 17; Barton 1999). The western boundary of the Sorell – Badger Head Zone is adjacent to an outcrop of blueschist facies rocks (Figures 14 box b, 15b; Calver & Reed 2001). The boundary between the Pedder and Tyennan zones, the east-dipping Mt Hobhouse Fault (Moore et al. 2012b) is seen in the offshore seismic data in southern Tasmania (Figure 16) (Drummond et al. 2000) and in the long wavelength magnetic data (Figure 1c). This fault has generally been placed along mapped faults; it also lies east of all of the ultramafic rocks in the Pedder Zone, placing it east of the Wings Subzone that has been inferred to have been emplaced westwards (Crawford & Berry 1992). It is also interpreted to lie adjacent to the only eclogite facies metamorphic rocks of western Tasmania (Figures 14 box a, 15a).
Cores in zircons from deformed granites yielded dates of 1051 ± 8 Ma and 1042 ± 35 Ma, and metamorphic monazites gave dates between 1015 ± 24 Ma and 920 ± 7 Ma (Berry et al. 2008). An undeformed syenite in the southwestern Glomar Zone yields a U-Pb SHRIMP date of 1119 ± 9 Ma and a TND age of 1270 Ma. Originally this location had been considered as part of the west South Tasman Rise, but the bathymetric, gravity and geological data (Figure 18) from the area have more in common with other parts of the east South Tasman Rise. All of these ages are unlike any from mainland Tasmania. As a result, we have chosen to regard the Glomar Zone as a separate geological entity.

Samples from the East Tasman Rise include granite gneiss, quartzite and marble with blue amphibole, all rocks unknown in eastern Tasmania, but the first two are also present in western Tasmania. No age date data are available.

The west South Tasman Rise is bathymetrically rough and typically lies in deeper water, mostly between 2000 and 4000 m. It also has a 200 μm^-1 lower gravity response. In the southern part, linear gravity anomalies generally trend approximately 010°, but in the north, anomaly trends vary between 070° and 140°. Detrital zircons in a dredged metasedimentary sample suggests that at least part of the west South Tasman Rise can be correlated with the Delamerian Orogen of southeastern mainland Australia (Berry et al. 2007), suggesting that the west South Tasman Rise may be better excluded from VanDieland. An alternative interpretation is that the west South Tasman Rise is similar to the Kanmantoo Group in South Australia or the Robertson Bay Terrane in north Victoria Land, both of which have Proterozoic deep crust or mantle at depth (Handler et al. 1997; Fioretti et al. 2005b; Rawlinson et al. 2014).

DISCUSSION

Tasmanian issues

Despite the apparent structural, metamorphic and stratigraphic complexity outlined above, there are many geological features that are shared across many of the tectonic zones (Figure 2). For example, the oldest exposed rocks are generally quartzites or quartzite-rich sequences with a minor but persistent detrital zircon population at
1450 to 1380 Ma. These are generally marginal marine facies, suggesting they were deposited on an older cryptic basement. Neoproterozoic extension occurred between ca 780 Ma and 580 Ma and is characterised by rifting, sedimentation and intermittent granite production. During this period western Tasmania was disaggregated into smaller crustal blocks separated by MORB and mafic–ultramafic complexes. The rifted blocks were deformed in the Tyennan Orogeny. Early west- or south-directed movement was accompanied by regional metamorphism, including blueschist facies in the Arthur Metamorphic Complex and eclogite facies along the extension to the Mt Hobhouse Fault (Figures 14, 15a, 16). The Mount Read Volcanics and the Owen Conglomerate were deposited during the waning stages of the Tyennan Orogeny across several tectonic zones, suggesting western Tasmania had reconfigured into a single crustal block. Subsequent deformation in the Tabberabberan Orogeny modified the boundaries and internal structures, but pre-existing structures can still be recognised.

Despite these similarities, there remain other observations that need to be addressed in order to develop a comprehensive tectonic model of VanDieland. These are discussed below.

Figure 16 Seismic line AGSO 148/15, upper panel showing interpretation based in part on that by Drummond et al. (2000), lower panel uninterpreted line, inset gives the line location. Shot points are approximately 50 m apart. The Mt Hobhouse Fault forms the boundary between the Pedder and Tyennan zones.
WHAT DO THE BOUNDARIES REFLECT?

Between the tectonic zones, most of the boundaries are defined by fault zones. However, several boundaries, such as the Tamar ‘Lineament’ and the Arthur Metamorphic Complex, are more distributed fault networks. Several of the boundaries may separate tectonic zones that are completely allochthonous with respect to their neighbouring zones (e.g. King Island Zone). Other boundaries may represent the inverted remnants of variably thinned crust reflecting that, at the end of the Proterozoic, western Tasmania was a series of crustal scale megaboudins (e.g. Péron-Pinvidic & Manatschal 2010; Direen et al. 2012).

The diversity in character and length of these boundaries suggests that different processes were active at the same time along the boundaries. For example, the high-pressure (to 1.9 GPa) metamorphism seen in the northern part of the Pedder–Tyennan boundary is consistent with craton–craton collision. By contrast, the southeastern segment of this boundary is characterised by lower metamorphic grade Proterozoic rocks (Calver et al. 2006) that do not indicate collisional tectonic processes. We suggest the Neoproterozoic to early Cambrian rifting of the Pedder and Tyennan zones resulted in ocean crust formation in the northern segment of the fault zone, while the southern zone was characterised by thin continental crust. We suggest a similar setting along the Rocky Cape–Burnie zone boundary, where the Arthur Complex has attained blueschist facies metamorphic grades in the northern part (Turner & Bottrill 2001), but the equivalent boundary to the southern is defined by mafic–ultramafic units that underwent lower grade metamorphism (Corbett 2003).

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**Figure 17** West end of TASGO Seismic Line 4, northern Tasmania, showing the boundaries of the Tyennan, Sorell–Badger Head and Eastern Tasmanian zones. The ramp anticlines in Eastern Tasmanian Zone may reflect mafic volcanic packages. Older west-dipping faults have been truncated by younger (Tabberabberan) faults. Dashed lines mark interpreted faults, solid lines marker horizons. No vertical exaggeration. Insert shows the location of imaged data. Shot points are approximately 50 m apart.
ARTHUR METAMORPHIC COMPLEX

The Arthur Metamorphic Complex (Arthur Lineament) separates mostly shallow marine Proterozoic rocks of the Rocky Cape Zone to the west from largely deep water Proterozoic metasediments of the Burnie Zone to the east (Seymour & Calver 1995). Holm & Berry (2002) mapped the Arthur Metamorphic Complex on both the western and northern Tasmanian coasts, and we interpret the boundary northward almost to the Victorian coast, and southward across the Sorell Peninsula (Figures 1, 3, 4). Holm et al. (2003) divided the rocks in the zone into three packages. The westernmost package was similar to rocks in the adjacent Smithton Basin. The central package comprised the rift-related mafic schist, amphibolite and metagabbro of the Bowry Formation, which was intruded by ca 777 ± 7 Ma granite (Turner et al. 1998) and metamorphosed to blueschist facies (350°C and 700 MPa; Turner & Bottrill 2001). The eastern package comprised rocks similar to the older parts of the Burnie Zone.

Geophysical modelling of the Arthur Metamorphic Complex suggests that none of the outcropping magnetic units persist to depths of more than 2 km (Leaman & Webster 2002). An east-dipping listric fault truncates the western edge of the package (Leaman & Webster 2002). A 10–15 km-wide body of weakly magnetic, dense rocks (p = 2.80 to 2.86 t m⁻³) is present under the Arthur Metamorphic Complex, suggesting the presence of either mafic igneous or high-grade metamorphic rocks. Further south, deep seismic data are consistent with the western boundary of the Arthur Metamorphic Complex having a dip to the east of approximately 30° and continuing to at least 5 s TWT, about 15 km depth (Figure 7; Leaman & Webster 2002).
Approximately 7 km offshore from the west Tasmanian coast, the Arthur Metamorphic Complex is displaced almost 50 km south (sinistrally) by the Braddon River Fault (Figures 1, 8). On the Sorell Peninsula, the boundary between the Rocky Cape and Burnie zones is an east-dipping thrust fault (Figure 8).

The dominant Cambrian movement sense in the Arthur Metamorphic Complex is of southwest-directed (sinistral) movement, with a later west-verging thrusting event (Holm & Berry 2002). Both events took place after the obduction of the mafic–ultramafic complexes at about 516 Ma. We interpret the Arthur Complex as crust that was extended at ca 770 Ma and then shortened in the Tyennan Orogeny and further modified in the Tabberabberan Orogeny. An implication of this is that the western part of the Burnie Zone may be underlain by equivalents to the older rocks in the Rocky Cape Block.

ORIGIN OF THE MAFIC–ULTRAMAFIC COMPLEXES

The understanding that VanDieland was a completely separate crustal fragment in the Cambrian (Cayley et al. 2011) has resolved the paradox of west-dipping subduction zones in Victoria (Miller et al. 2005) and Antarctica (Federico et al. 2006) lying either side of an east-dipping subduction zone in Tasmania (Crawford & Berry 1992; Moore et al. 2012a). However, the question remains as to where the mafic–ultramafic complexes were obducted from.

Crawford & Berry (1992) considered that they came from an unseen but speculated Cambrian arc system that lay to the east of VanDieland. In this interpretation, mafic–ultramafic units were obducted at least 130 km westwards over the Tyennan and Pedder zones while leaving no evidence for their occurrence to the east, closer to their interpreted origin. We pose an alternative model, in which the mafic and ultramafic rocks with oceanic and transitional crust affinities were formed between micro-continental fragments that separated during 250 Ma of Neoproterozoic extension. In the Tyennan Orogeny, these basins were inverted and the individual ribbons re-amalgamated to produce the geometries and lithological distributions in western Tasmania. The model does not require an arc system as proposed by Crawford & Berry (1992). Obducted mafic–ultramafic units in the Burnie Zone were derived from the inversion of thinned continental crust (perhaps with some oceanic crust) along the deep, east-dipping Penguin Fault (Rawlinson & Urvoy 2006; Rawlinson et al. 2010) that forms part of the western boundary of the Pedder and Tyennan zones (Figures 1, 11, 12a). This model requires obducted slices to travel ~30 km, not 130 km. Inversion of small transitional crust to oceanic back arc basins may not have initiated subduction, as inversion may have been accommodated by subduction and thrust slicing of the transitional and ocean crust substrate in a manner similar to that imaged in Bendigo Zone of the Lachlan Fold Belt (Cayley et al. 2011). The driver for this inversion may have been a convergent margin located outboard of Tasmania. A similar scenario might have seen the mafic–ultramafic complex rocks in the Tyennan Zone obducted west from the western boundary of the Sorell–Badger Head Zone.

An east-dipping mafic–ultramafic complex lies close to the western edge of Seymour & Calver’s (1995) Adamsfield–Jubilee Element (Figure 1a). Mapping implies that another significant boundary lies west of the mafic–ultramafic complex (Calver et al. 2006). We interpret the mafic–ultramafic rocks and the rocks either side to have been within a leading-edge thrust system, with the master fault being the Mt Hobhouse Fault (Moore et al. 2012b). While it is somewhat arbitrary as to whether one considers the Adamsfield–Jubilee area part of the Tyennan Zone or the Pedder Zone, we have chosen the latter in order to emphasise the significance of the Mt Hobhouse Fault.

MOUNT READ VOLCANICS

The Mount Read Volcanics have been interpreted as a volcanic arc that formed above a west-dipping subduction zone during the late Cambrian (Crawford & Berry 1992). In this model, a transient, west-dipping subduction zone formed to the east of the ‘Tamar Lineament.’ However, there seems to be little independent evidence for this subduction zone. Recent geochronology results by McNeil et al. (2012) from the Mount Read Volcanics demonstrate that they were formed over 12 m.y., apparently longer than the ‘short-lived’ event envisioned by Crawford & Berry (1992), and potentially questioning the validity of their interpretation. Meffre et al. (2004) described an east-dipping sequence at least 15 km thick of a rift-related tholeiitic volcanic passive margin sequence that crops out on King Island. This magnetic sequence can be traced southwards approximately 50 km to the west of the Mount Read Volcanics (Figures 1, 3). Inversion of this passive margin is likely to have resulted in the amalgamation of the King Island Zone with the rest of western Tasmania, and as this took place, the Mount Read Volcanics formed as the east-dipping subducting slab delaminated. We consider that this is a more satisfactory solution to the origin of the Mount Read Volcanics than that proposed by Crawford & Berry (1992).

WICKHAM OROGENY

The Tasmanian Cryogenian and Ediacaran geological record is dominated by events that can be interpreted as extensional (Figure 2; Li 2001). The Wickham Orogeny was identified on the northern and western parts of King Island, where granite intrusion and associated widespread contact metamorphism and tholeiitic dyke emplacement took place at $760 \pm 12$ Ma (Black et al. 1997; Calver 2004; Calver & Everard 2014), consistent with extension. These rocks are coeval with a $777 \pm 7$ Ma granite and rift-related tholeiites in the Arthur Metamorphic Complex (Turner et al. 1998; Holm et al. 2003). Everard et al. (2007) showed that deposition occurred into the extensional Smithton Basin from the middle Cryogenian (ca 750 Ma). We therefore consider that the Wickham Orogeny is likely to represent a period of crustal extension, rather than one of crustal shortening, and is likely to record Rodinia break-up in western Tasmania (Holm et al. 2003).
PRE-GONDWANA BREAKUP LOCATION OF THE GLOMAR ZONE

There is no consensus to the precise pre-breakup alignment of Australia and Antarctica. Several authors have favoured the alignment of the Darling Fault in Western Australia with Denman Glacier Antarctica (Fitzsimons 2003; Finn et al. 2006; Goodge et al. 2010), whereas others have preferred a closer alignment of the eastern edge of the buried Gawler Craton with the western edge of Ross Orogen (Fioretti et al. 2005b; Boger 2011; Lisker & Läufer 2013), or intermediate locations (Williams et al. 2012; Aitken et al. 2014). Placing smaller crustal fragments like the Glomar Zone in the context of Rodinia and Gondwana is even more subjective because their origins and original geometries are poorly constrained, and so there is no consensus about the original position of the Glomar Zone. Norvick & Smith (2001) placed it to the southeast of Tasmania, while Williams et al. (2011) showed it to the west at about 160 Ma, and Royer & Rollet (1997) had it in its present location at 95 Ma. The reconstruction of Norvick & Smith (2001) seems unlikely, as it places +1050 Ma crust east of our interpreted position of the eastern margin of VanDieland Proterozoic crust. The reconstruction by Williams et al. (2012) appears to be equally unlikely, since it places the Glomar Zone, with granites with ages of 1050 Ma and metamorphic rocks at 920 Ma, adjacent to King Island, with metamorphism at 1290 Ma and granites at approximately 760 Ma, and there are no known thermal events between 1050 and 920 Ma on King Island. We therefore consider the most likely location for the Glomar Zone before Gondwana breakup to be close to its present position south of Tasmania, as originally suggested by Royer & Rollet (1997). This interpretation may be supported by the presence of Cretaceous northwest-trending basins on the east South Tasman Rise (Exon et al. 1997b), which are parallel with the dominant trends in the free-air gravity anomalies from with the Glomar Zone (Figure 18). Furthermore, K/Ar ages determined on biotite in the Glomar Zone gave ages of 511 ± 4 Ma and 451 ± 4 Ma (Berry et al. 1997), consistent with Tyennan metamorphism. These data suggest there was not sufficient crustal extension or increase in the geothermal gradient to exceed the closure temperature of biotite, approximately 300°C, during Gondwana break-up.

Possible correlations with VanDieland

A fundamental part of deriving a coherent geological history for VanDieland is to place it within the wider context of Earth history. The Proterozoic events seen in VanDieland took place within the assembly and dispersal of Rodinia and the formation of Gondwana (Calver et al. 2014), and so events elsewhere should be reflected in the history of VanDieland. Because there is only one Proterozoic paleomagnetic data point of doubtful reliability from VanDieland (McWilliams & Schmidt 2003), other data assume greater importance in determining the most likely relationships with other cratons. We accept the reservation outlined by Andersen (2014) that identical detrital zircon populations can form at the same time on cratons distant from each other. Nevertheless, even though these data are non-unique, they can exclude some possibilities. As well, where other data are available, we have attempted to integrate them as closely as possible into the discussion and Table 2.

ELSEWHERE IN AUSTRALIA

The provenance of the detrital zircons in Tasmania provides important constraints on where any Tasmanian crustal fragments may have come from. The dominant population, from approximately 1900 to 1650 Ma, is widespread in Australia and may have been derived from any of the Nuna assembly events (Evans 2013). In contrast, the 1460 to 1380 Ma zircons seen in VanDieland are less common elsewhere (Condie et al. 2009) and so provide a tighter constraint. In Australia, the only possibilities appear to be the Western Australian Madura Complex (near the South Australian border and covered by Cenozoic sedimentary rocks), which is known to include meta-igneous rocks containing zircons with the appropriate ages (Nelson 2006a, b), or the Musgrave Province in central Australia (Kirkland et al. 2013). However, if VanDieland has always been in generally the same position relative to the rest of Australia, as suggested in most Rodinian assemblies (e.g. Li et al. 2008b), the drainage required to transport the 1400 Ma zircons must have crossed the Mawson Craton, which includes the 1600 to 1500 Ma Hiltaba, Spilsby and St Peter suites, but there is little evidence of these older zircons in the VanDieland meta-sedimentary rocks. In contrast, 1600 to 1500 Ma zircons are abundant in the Neoproterozoic Adelaide Geosyncline metasedimentary rocks, which were also marginal to the Mawson Craton (Ireland et al. 1998; Preiss 2000).

An alternative interpretation is that VanDieland was once adjacent to the Madura Province in southeasternmost Western Australia. However, the paleomagnetic evidence suggests that the Northern Australian, Western Australian and Mawson cratons have either maintained their present relative positions since 1500 Ma (Wingate & Evans 2003), or at most rotated approximately 40° between 650 and 550 Ma (Li & Evans 2011). Although this interpretation can be used to correlate the 1290 Ma metamorphism seen on King Island and the Rocky Cape Zone (Berry et al. 2005; Halpin et al. 2014) with that seen in the Albany-Fraser Orogen (Bodorkos & Clark 2004), it raises other problems such as extracting VanDieland and then positioning the other cratons back into their previous locations. Detrital zircons found in the Anakie Inlier in central Queensland include a significant 1300 to 1100 Ma population (Fergusson et al. 2001), similar to those seen in the Sorell—Badger Head Zone and are coeval with the 1290 Ma metamorphism on King Island, but other populations typical of those found on King Island or in the Sorell—Badger Head Zone are not present, suggesting the two regions have little in common. The presence of detrital 1370 Ma monazite in VanDieland, an age absent from Australian terranes, adds yet another layer of difficulty in locating a possible Australian source area.

A further difficulty is that the Mesoproterozoic and Neoproterozoic history of VanDieland seems a poor fit with other regions in Australia. There are many areas
along the eastern Rodinian breakup margin where 1900 to 1600 Ma basement is present. However, none seem capable of generating the quartzites derived from what appears to have been a Grenville-aged source; all possibilities are in Western Australia or have been previously eliminated by the problem of the 1400 Ma zircons (Betts et al. 2002).

LAURENTIA

We consider that the southwestern margin of Laurentia is a potential source area for much of the detritus in western Tasmania. The clean quartzites of both southwestern Laurentia and western Tasmania and equivalents contain a dominant detrital zircon population of 1800 to 1700 Ma, and traces of >2400 Ma zircons (Figure 19) (Black et al. 2004; Jones et al. 2009). As well, metasedimentary rocks from both regions host minor detrital zircon populations between 1450 and 1380 Ma (Black et al. 2004; Jones et al. 2011). Southwestern Laurentia was also a potential source of 1100 Ma igneous zircons (Bright et al. 2014) that are present in many of the western Tasmanian detrital zircon populations (Black et al. 2004). Because of the multiple sources suggested, it is not possible to carry out a meaningful statistical analysis such as that outlined by Sircombe & Hazelton (2004). Metasedimentary rocks from central Tasmania also include 1370 Ma detrital monazites (Chmielowski 2009) that could reasonably have been derived from the western Laurentian Granite-Rhyolite Province and other A-type granites in the Yavapai and Mazatzal orogens (e.g. Jones et al. 2011).

The excess of ca 1600 Ma inherited zircon populations in the Tasmanian Devonian granites relative to the host Rocky Cape Group (Black et al. 2010) suggests the zircon source from an unexposed VanDieland Mesoproterozoic basement that lies beneath the Rocky Cape Group. We suggest that this basement may correlate with the Yavapai-Mazatzal terranes of southwest Laurentia. Limited TDM model age data of 1700 Ma or younger are consistent with data from the Cochise Block in the Mazatzal Orogen (Whitmeyer & Karlstrom 2007).
VanDieland also shows evidence of Grenville-age events. Detrital zircon populations with ages ranging from 1268 ± 19 Ma to 914 ± 44 Ma (Black et al. 2004) are common, and the Glomar Zone includes granites with ages ranging from 1119 ± 9 Ma (Fioretti et al. 2005a) to 1042 ± 35 Ma, which were metamorphosed at 918 ± 9 Ma (Berry et al. 2008). We suggest that the Glomar Zone correlates with the Grenville Front in southwestern USA, and that the Grenville-Glomar zone zircons were subsequently eroded to be deposited in the upper Rocky Cape Group and the Badger Head Group. The ca 770 Ma Wickham Orogeny in VanDieland appears to be coincident with widespread rifting in Laurentia, in which the ca 800 to 740 Ma Chuar Group was deposited in Arizona (Timmons et al. 2001), and ca 780 Ma mafic intrusions were emplaced in a belt that extends from Montana to the Great Slave Lake (Harlan et al. 2003). The final breakup of the southwest Laurentian margin occurred at about 580 Ma (Yonkee et al. 2014), coeval with the 580 Ma break-up in VanDieland.

Halpin et al. (2014) suggested that the Rocky Cape Group in Tasmania could be correlated with the Belt-Purcell Supergroup in Laurentia. However, Medig et al. (2014) correlated the 1500 Ma Mt Isa granite event with detrital zircon populations found in the Coal Creek Inlier in the Yukon, and hence the Mawson Craton granites with input into the Belt-Purcell Supergroup. This correlation places potential Laurentia–East Antarctica correlates of VanDieland, much further to the south than the Belt-Purcell rocks. As well, the hypothesis of Halpin et al. (2014) seems to take insufficient account of the 1120 to 1040 Ma zircon igneous age dates (Fioretti et al. 2005a; Berry et al. 2008) and approximately 920 Ma metamorphic monazite dates (Berry et al. 2008) from granites from the Glomar Zone. Like Fioretti et al. (2005a), we prefer correlating the Glomar Zone with the Grenville Orogen of Laurentia.

EAST ANTARCTICA

There are few clear links between VanDieland and the pre-1300 Ma rocks of East Antarctica, including the Mawson Craton (Payne et al. 2009). However, the prominent 1900 to 1650 Ma detrital zircon peak in the Tasmanian quartzites overlaps with the 1730 to 1720 Ma Nimrod Orogeny in East Antarctica (Figure 19; Goodge et al. 2001). Clean metaquartzites in the area of the Princess Anne Glacier in the Queen Elizabeth Range also contain first cycle zircon populations with peaks between 1800 and 1700 Ma and lesser peaks from 1450 to 1400 Ma (Goodge et al. 2004), consistent with similar sources to the Tasmanian quartzites (Black et al. 2004). Although Goodge et al. (2004) did not record paleocurrent directions from the Antarctic quartzites, other units indicated sources from under the ice cap, and they considered that the quartzites had come from the same direction.

Goodge et al. (2008) recorded the presence of a 1441 ± 6 Ma (U–Pb zircon SHRIMP date) A-type granite clast in a moraine a few tens of kilometres north of the quartzites and indicated that it came from under the ice cap. This clast was correlated with the Granite-Rhyolite Province in Laurentia, suggesting southwestern Laurentia abutted the central Transantarctic Mountains during Rodinia assembly. Similar granites may have sourced both the 1450 to 1380 Ma detrital zircon and the 1370 Ma detrital monazite in the Tasmanian quartzites. Goodge et al. (2010) strengthened this Laurentia–East Antarctica correlation by finding meta-igneous clasts in moraines with protolith ages of 1100 to 1000 Ma. These ages lie between the 1290 and 760 Ma metamorphic ages seen on King Island (Turner et al. 1998; Berry et al. 2005), but are consistent with the 1250 to 1000 Ma detrital zircons seen in the Rocky Cape and Sorell–Badger Head zones (Black et al. 2004). They also accord with the 1120 to 1040 Ma
ages of dredged granite and orthogneiss from the Gl prah Zon (Fioretti et al. 2005a; Berry et al. 2008).

Borg & DePaolo (1991, 1994) used Sm–Nd isotopic data to propose a strip of Nimrod Group basement that extended outboard of southern Victoria Land through to the Central Transantarctic Mountains and as far south as the Beardmore Glacier. This outboard terrane is characterised by $T_D^{DM}$ model ages of 1900 to 1600 Ma ($e_Nd$ − 15.4 to −10.0 at 500 Ma), which are significantly younger than the $\gtrsim$2000 Ma model ages for the inboard Antarctic terranes. Recent data support this division (Peucat et al. 2002; Goodge et al. 2012). Ca 1600 Ma granites are present in the Mawson Craton under the Antarctic ice (Peucat et al. 2002; Betts et al. 2008), but it is unlikely that the middle and lower VanDieland crust can be correlated with the deep Antarctic crust, because the Antarctic model ages are older than 2000 Ma where these intrusions are present.

Other correlations can also be made. Goodge et al. (2002) recorded the presence of minor ca 770 and 765 Ma detrital zircon populations in the central Transantarctic Mountains, which may be derived from the same igneous event as ca 770 Ma rift-related granites in VanDieland. Rhyolitic clasts in the Skelton Group in southern Victoria Land yielded a 650 Ma age (Cooper et al. 2011), within error limits of a 660 Ma deformed granite near Badger Head in northern Tasmania (Black 2007, unpublished data). A similarly aged (668 ± 1 Ma, U–Pb zircon) gabbro was intruded into the middle of the Beardmore Group (Goodge et al. 2002), implying that rift-related sedimentation must have commenced earlier. The upper age of the Beardmore Group is poorly constrained, but an unconformably overlying package contained detrital zircons with ages as old as 589 ± 6 Ma, which Goodge et al. (2012) interpreted as indicating that arc volcanism had commenced by then. However, the age is indistinguishable from that of 582 ± 4 Ma zircons in a rift-related rhyolite within a basaltic sequence in the upper Togari Group in Tasmania (Calver et al. 2004), suggesting that similar rift-related rhyolites may have been the source to the 589 ± 6 Ma zircons. Lastly, it may be no coincidence that both the early Beardmore Group and the possibly coeval Onah Formation in Tasmania are quartz rich at the base, but become more carbonate rich near the top (Brown 1986; Goodge et al. 2002).

**CATHAYSIA**

Li et al. (2008b) and Li & Evans (2011) suggested that the Cathaysia Block lay between Laurentia and the Northern Australian Craton, which could suggest that VanDieland once lay adjacent to Cathaysia. This correlation is strengthened by the presence of similar zircon populations in the Sorell–Badger Head Zone to those from the Shihuiing Formation on Hainan Island at the southern end of the Cathaysia Block (Li et al. 2008a) (Figure 19); both regions have detrital zircons with age peaks at ca 1000 Ma, and 820 to 720 Ma igneous zircons present in the southeastern-most Cathaysia Block cover the age of the 770 to 740 Ma Wickham Orogeny (Black et al. 1997, 2004; Li et al. 2014).

Zheng et al. (2011) considered that most, if not all, of the Cathaysia Block was underlain by Archean crust. However, Li et al. (2014) suggested that these Archean ages were derived from far-travelled zircons, and that a more appropriate basement age is approximately 1850 Ma or younger. Regardless of which is correct, neither age supports a correlation with VanDieland, where the data suggest a $T_D^{DM}$ model age of approximately 1700 Ma or younger. Furthermore, Cawood et al. (2013) considered that the presence of 1000 to 900 Ma arc and back-arc rocks along the western margin of the Cathaysia Block indicated it must have been located on the edge of Rodinia, perhaps off the Western Australian Craton, while Li et al. (2014) considered that these ages reflected plume-related breakup. Both the nature of the southernmost Cathaysia Block on Hainan Island and whether these volcanic rocks are arc-related or break-up plume-related are beyond the scope of this paper. Thus, we consider that it is possible that VanDieland and southeastern-most Cathaysia might be able to be correlated, but that there are too many uncertainties to be more definite.

**BALTICA**

In his ANACONDA model of Rodinia, Evans (2009) placed Baltica between Laurentia and the Southern Australia–Mawsonland craton. Zircons of the appropriate 1450 to 1380 Ma age were produced in the Danopologian Orogeny of southwestern Baltica, suggesting that a correlation with VanDieland might be possible. As well, widespread remnants of older quartzites are present throughout much of Baltica (Bogdanova et al. 2008), suggesting a possible source for the quartzites in VanDieland. However, other factors seem incompatible with this interpretation. Bogdanova et al. (2008) described a semi-continuous sequence of zircon-producing igneous and metamorphic events that lasted from 1730 to 1480 Ma. This contrasts with the detrital zircon patterns in VanDieland, which typically show a population gap between approximately 1650 and 1450 Ma (Black et al. 2004), and seems less likely to provide the contrast in inheritance between the Paleozoic granites and the Proterozoic metasediments (Rocky Cape Zone; Black et al. 2010). Finally, we note the slightly different breakup ages between Fennoscandia, 800 Ma to 550 Ma (Andréasson et al. 1998), and VanDieland, 770 Ma to 575 Ma. While not of themselves critical differences, they too suggest that other hypotheses may be stronger.

**SYNTHESIS, CORRELATIONS AND CONSTRAINTS: A REGIONAL TECTONIC HISTORY**

This section connects the previous interpretations and links them with others along the Australian and Antarctic Rodinia to Gondwana margin.

The xenocrystic zircons in the Paleozoic granites in western Tasmania indicate the presence of an older, unexposed substrate with a strong 1650 to 1600 Ma igneous component. We suggest, first, that the source of these zircons is the Mazatzal Orogen of the Laurentian craton, which Goodge et al. (2010) indicated was adjacent to East Antarctica from at least 1400 Ma to ca 800 Ma, and second, that VanDieland lay between the two (Figures 20, 21). An implication is that we favour a broadly SWEAT-
Figure 20 Geological history of VanDieland.
like reconstruction of Rodinia for Australia, Antarctica and Laurentia similar to that suggested by Goodge et al. (2010), Cawood et al. (2013) or Li et al. (2013).

The oldest rocks outcropping in the Tasmanian region are at the base of the Rocky Cape Group and on King Island, and were metamorphosed at ca 1300 Ma (Berry et al. 2005; Halpin et al. 2014). If a Rodinian fit close to those suggested by Goodge et al. (2010) or Li & Evans (2011) is valid, then it appears likely that the 1300 Ma metamorphism is connected to the Grenvillian events in southwestern USA or East Antarctica. We also suggest that the Grenvillian Front can be mapped at the northern edge of the Glomar Zone (Figure 18), since Grenvillian-age rocks have been recorded on the Glomar Zone but not in Tasmania.

As Rodinia began to break up, at approximately 830 Ma (Wingate et al. 1998), extension took place along the proto-Terra Australis margin. In VanDieland, it was marked by the 780 to 750 Ma Wickham Orogeny and sedimentation beginning in the Smithton Basin, coeval with the 777 ± 7 Ma bimodal Boucaut Volcanics in the Adelaide Geosyncline (C.M. Fanning personal communication, 1994, in Preiss 2000). Continued extension is indicated by the intrusion of ca 710 Ma rift tholeiites in the Burnie Formation and also by the beginning of sedimentation in the Beardmore Group in the Transantarctic Mountains, which pre-dates 668 ± 1 Ma pillow basalts and gabbro there (Goodge et al. 2002, 2004). Sedimentation also commenced at about 770 Ma in southwestern North America (Timmons et al. 2001; Yonkee et al. 2014). The 660 Ma intrusive age of the deformed granite in the Sorell–Badger Head Zone is also close to the age of the gabbro in the Beardmore Group, suggesting that they may have formed during the same extensional event. We interpret this rifting to have extended the Pedder, Tyennan and Sorell–Badger Head zones into crustal-scale megaboudins (Figure 20).

After ca 650 Ma, strain became more distributed, allowing continued deposition in the Smithton Basin and equivalents, but perhaps also the Success Creek Group, which unconformably overlies the Oonah Formation (Brown 1986). Rifting also took place elsewhere along the Rodinian margin. In western Victoria, Neoproterozoic metasedimentary rocks were intruded by gabbro dykes that yielded primary zircons with an age of 643 ± 4 Ma (Morand & Fanning 2006). In the Transantarctic Mountains, deposition continued in the Beardmore Group (Goodge et al. 2004). A seaway had developed along the entire western USA (Yonkee et al. 2014). The final rifting event in VanDieland is marked by the 575 ± 3 Ma MORB tholeiitic basalts of the Grassy Group on King Island (Calver et al. 2004; Meffre et al. 2004) and the 582 ± 4 Ma rift tholeiitic basalt of the Spinks Creek Volcanics in the Smithton Basin (Calver et al. 2004). The Crimson Creek Formation of the Burnie Zone and other mafic volcanioclastic packages in western Tasmania were probably deposited at this time or shortly before (Seymour & Calver 1998). These events left the King Island Zone isolated from both East Gondwana and VanDieland, while the Rocky Cape and Burnie zones remained contiguous. The Tyennan and Pedder zones were separated at their (now) northern ends, but connected by subcontinental lithosphere further south, similar to the present western North Atlantic Ocean, where the Hatton, Rockall and Porcupine banks are partly separated by V-shaped basins (Peron-Pinvidic & Manatschal 2010). The Tyennan and Sorell–Badger Head zones were also contiguous. This event slightly post-dates the last rifting event in the western Lachlan Fold Belt (586 ± 3 Ma; Greenfield et al. 2011), which is evidenced by ca 580 Ma detrital zircon population in the Delamerian Orogen (Ireland et al. 1998; Fanning & Morand 2002; Morand & Fanning 2009). In East Antarctica, 589 ± 6 Ma zircon cores surrounded by 559 ± 6 Ma rims in a clast of muscovite–biotite granite may record the same event (Goodge et al. 2012). The rims and the younger of the locally derived 580 to 540 Ma detrital zircon populations seen in the Cambrian Byrd Group in the Transantarctic Mountains (Goodge et al. 2002) may mark the transition to a convergent margin. Breakup was also completed in Laurentia (Yonkee et al. 2014), suggesting that VanDieland was left as an isolated microcontinental ribbon between the larger Gondwanan and Laurentian cratons.

After this, VanDieland drifted as at least two and perhaps as many as seven fragments on the proto-Pacific Ocean. In Tasmania, upper Neoproterozoic sedimentation was restricted to carbonate deposition. Lower Cambrian laminated siliceous siltstone and shale deposits suggest VanDieland was isolated from clastic sources (Burns 1964; Everard et al. 2007). The rifted fragments of VanDieland migrated northwards along the earliest Gondwanan margin and were not affected by the ca
560–530 Ma early phases of the Ross Orogeny (Allibone & Wysoczanski 2002; Stump et al. 2006; Goodge et al. 2012). Following a period of relative tectonic quiescence, the separated fragments of VanDieland began to converge at ca 530 Ma (Figures 20, 22), perhaps beginning to generate the high-pressure Franklin Metamorphic Complex at 529 ± 10 Ma, before they underwent decompression at 512 ± 4 Ma (Chmielowski 2009; Chmielowski & Berry 2012). Convergence between the Pedder and Tyennan zones occurred along the Mt Hobhouse Fault and may have coincided with the westward obduction of the mafic–ultramafic complexes in the region at 516 ± 1 Ma (Mortensen et al. in press). Final docking of these zones post-dated movement along the Mt Hobhouse Fault because the eastern edge of the Burnie Zone truncates the fault. During this event, the Wings Subzone in the Adamsfield area was trapped and obducted westwards onto the Pedder Zone. At a regional scale, these events are coeval with the 514 ± 4 Ma U/Pb zircon age for the syn-tectonic Rathjen Gneiss in South Australia (Foden et al. 1999). Deformation associated with the Ross Orogeny also continued in Antarctica (Goodge et al. 1993) and was associated with felsic magmatism (Encarnación & Grunow 1996), associated with west-dipping subduction in the region (Boger & Miller 2004).

By 510 Ma, collisional events were close to peaking in Tasmania. They drove the Tyennan metamorphism and deformation, including in the basement in the Sorell–Badger Head Zone in the east (monazite, 513 ± 8 Ma; Chmielowski 2009), in the Forth Metamorphics in the central north of the Tyennan Zone (zircon, 512 ± 5 Ma, Black et al. 1997 recalculated by Foster et al. 2005; monazite, 510 ± 11, Chmielowski 2009), and in the Pedder Zone in the southwest (e.g. monazite, 511 ± 5 Ma, Chmielowski 2009). Inversion and cooling started along the Mt Hobhouse Fault (monazite, 512 ± 4 Ma, Chmielowski & Berry 2012). Holm & Berry (2002) described north–south shortening along the Arthur Metamorphic Complex in the west. In the east, northeast–southwest shortening took place, thrusting the Badger Head Block and a west-dipping mafic–ultramafic slice over the Port Sorell Formation (D2 in Reed et al. 2002). If the two shortening events were coeval, they would be likely to have driven some mafic–ultramafic complexes south, as described by Berry & Bull (2012). More regionally, Preiss (2001) interpreted an initial northwest–southeast plate convergence at about 510 Ma. This was followed by northward propagating transpression. In the Koonenberry Block in western New South Wales, northeast–southwest extension was accompanied by basaltic to rhyolitic volcanism (Greenfield et al. 2011). Deformation and magmatic activity also continued along the East Gondwanan Antarctic margin (Cawood & Buchan 2007, and references therein).

The Mount Read Volcanics began erupting at 506.8 ± 1.0 Ma (U/Pb, zircon; McNeill et al. 2012; Mortensen et al. in press). We attribute this to subduction associated with the arrival and collision of the King Island Zone. We note that probable blueschist metamorphism in southwestern Tasmania continued until 505 ± 2 Ma and the white-schist metamorphism along the Mt Hobhouse Fault at 504 ± 5 Ma (U/Th/Pb, monazite; Chmielowski & Berry 2012), suggesting shortening was maintained until this time. The Mount Read Volcanics continued to erupt.

Figure 22 Reconstruction of western Tasmanian assembly. Compare with Figures 1 and 20. At 560 Ma, VanDieland was a series of cratonic blocks, either loosely connected by transitional crust, or completely separated as in the case of the King Island Zone. By 510 Ma or shortly thereafter, the Tyennan Orogeny had brought most of the cratonic blocks together. By 485 Ma, the remaining blocks had been accreted, and most of the movement along the Braddon River Fault had taken place.
until 496 ± 0.9 Ma (Mortensen et al. in press), coeval with the last thermal event in north central Tasmania at 497 ± 3 Ma (U/Th/Pb, monazite; Chmielowski 2009). We suggest that the volcanism and shortening were due to an east-dipping subduction zone between King Island and Tasmania that rolled back after about 505 Ma.

The Mount Read Volcanics are contemporaneous with the geochemically similar Stavely Volcanic Belt in western Victoria (Crawford et al. 2003a). Volcanic rocks from the Stavely Volcanic Belt yield U–Pb SHRIMP dates of 501 ± 9 Ma and 495 ± 5 Ma (Stuart-Smith & Black 1999), similar to Delamerian magmatism in South Australia (Foden et al. 2006) and western New South Wales (Greenfield et al. 2011). This magmatism occurred in the overriding plate of a west-dipping subduction zone along the main East Gondwana margin, inboard of Tasmania (Foden et al. 2006; Greenfield et al. 2011). The cessation of Tasmanian volcanism is coincident with movement on the Moyston Fault in western Victoria at ca 495 Ma (40Ar/39Ar on mica and hornblende; Miller et al. 2005). This fault has a major dip-slip component, with the hangingwall of amphibolite facies rocks and footwall of prehnite–pumpellyte facies rocks (Cayley & Taylor 2001) and is interpreted to represent the relics of a west-dipping subduction zone (Miller et al. 2005). However, the fault also has a significant component of sinistral strike-slip movement, suggesting that oblique convergence in the region would have been likely (Cayley & Taylor 2001). The movement sense is consistent with the regional kinematics in the back-arc region in South Australia (Preiss 2001), suggesting northwest oblique convergence outboard of the Cambrian East Gondwana margin continued for at least 15 Ma. It is also coeval with subduction associated with the Narooma Complex on the south coast of New South Wales (Prendergast & Offer 2012). In the same period, our model suggests that the individual tectonic elements that comprise VanDieland were accreting to the west of the Narooma Complex (Moore et al. 2012a) but outboard of the Gondwana margin.

After these accretion events, at the end of the Cambrian and into the Early Ordovician, western Tasmania underwent extension, perhaps caused by further rollback and/or by retreat of the eastern subduction zone. As a result, the locally derived Owen Conglomerate was deposited in half-graben on the west coast and in the north as far east as the eastern part of the Sorell–Badger Head Zone (Noll & Hall 2005; Reed & Vicary 2005). Similar aged, locally derived conglomerate is also present in western New South Wales (Greenfield et al. 2011) and probably also in north Victoria Land (Tessensohn & Henjes-Kunst 2005). Near VanDieland, subduction stalled, perhaps as a result of the collision of the Dimboola Complex into the early Gondwana margin (Moore 2006). In Antarctica, granite intrusion and deformation continued into the Ordovician, until at least 480 Ma (Goodge et al. 2004; Rossetti et al. 2011). At least one granite, the 512 ± 3 Ma Surgeon Island Granite, shows strong affinity in its inherited zircon population with granites in western Tasmania (Fioretti et al. 2005b). As well, the eastern Robertson Bay Terrane appears to be underlain by a different basement from the rest of north Victoria Land (Fioretti et al. 2005b). This suggests that VanDieland and the eastern Robertson Bay Terrane, separated in the Gondwanan breakup, were once part of the same microcontinental fragment. Thus, the terrestrial source of the Cambrian sedimentary rocks of the Robertson Bay Terrane (Henjes-Kunst & Schüssler 2003; Tessensohn & Henjes-Kunst 2005) may have been the erosional products of the Tyennan Orogeny in VanDieland.

At this time, the only paleopole available from northwestern Tasmania places it close to the East Gondwana coast, if not abutting it (Li et al. 1997). Evidence from Victoria suggests that the former is more likely. During the Late Ordovician to Silurian Benambran Orogeny, the Bendigo Zone was shortened by ~200 km (Gray et al. 2006; Cayley et al. 2011), while the western Melbourne Zone was shortened by ~40 km in the Middle Devonian Tabberabberan Orogeny (Foster & Gray 2007), which was accommodated by the Bendigo Zone over-riding the Selwyn Block (Cayley et al. 2011). This shortening is well within the most likely error limits given by Li et al. (1997), α95 = 10.4° (i.e. ~700 km east–west and ~1200 km north–south).

Our accretion model for VanDieland requires broadly similar plate kinematics to that defined on the late Cambrian eastern margin of Gondwana (e.g. Cayley & Taylor 2001; Cayley 2011). The horizontal accretion of micro-continental slivers or ribbons took place when a central core comprising the Tyennan and Pedder zones successively accreted or obducted the Burnie–Arthur–Rocky Cape, Sorell–Badger Head and King Island zones during north-to-northwest relative movement (Figure 22). During this accretion, north–south shortening took place in the Trial Harbour area on the west coast (McFarlane 2011), in central Tasmania (Calver et al. 2006) and the Arthur Lineament (Holm & Berry 2002). Our model suggests that the internal collisions were oblique, consistent with the kinematic observations. While the magnetic and gravity interpretation provides a clear sequence of accretion events, the polarities of the subduction zones between the accreted fragments are largely unconstrained. Exceptions are the collision between the combined Tyennan and Pedder zones and the Burnie Zone, where westward obducted mafic–ultramafic complexes indicate east-dipping subduction (Crawford & Berry 1992), and the King Island–western Tasmanian subduction zone, which is also east dipping. Cayley (2011) proposed that VanDieland briefly accreted to the eastern margin of Gondwana at about 500 Ma. This is tentatively supported by trilobite fauna (Hally & Paterson 2014), which shows a convergence of shallow-water species in Gondwana and western Tasmania at that time.

At the global scale, these later events lie between the early Gondwana continent and the great circum-Pacific subduction system that has existed throughout the Phanerozoic (e.g. Cony 1992; Cawood 2005). We suggest that at least two Tasmanian fragments broke off in the dispersal of Rodinia at about 580 Ma (Yonkee et al. 2014) and these were left isolated in the newly formed ocean. Whether this was due to rifting or subduction roll back that caused back arc extension in the overriding plate is not clear. After VanDieland had departed, East Antarctica began to accrete adjacent terranes to form Gondwana (Boger 2011). However, in the region of the Transantarctic Mountains, the Neoproterozoic continental fragments remained stranded in a back-arc setting.
between a retreating west-dipping subduction zone and the Gondwanan margin. VanDieland was left outboard of the subduction system that started in the Antarctic region at about 560 Ma (Goodge et al. 2012) and the larger subduction system that developed in the Cambrian along the early Gondwanan margin (Cawood 2005; Casquet et al. 2012). During inversion of this back arc setting the micro-continental fragments and megaboudins of VanDieland migrated northwards, and from ca 520 Ma to ca 495 Ma they successively accreted within this closing back arc system. Unlike the Beardmore and Bowers terranes along the Antarctic margin (Stump et al. 2006; Godard & Palmeri 2013), VanDieland did not accrete back onto the early Gondwanan margin at this time. Rather, it moved closer in the Early Silurian and, together with eastern Tasmania and the Lachlan Orogen, was finally integrated with the Gondwanan craton in the Middle Devonian.

The accretion of different fragments in different ways along the Gondwanan margin in the Ross–Tyennan–Delamerian Orogeny would inevitably lead to different effects at different times along the same margin. The timings range from 554 ± 10 Ma (\(^{87}\)Rb/\(^{86}\)Sr) in South Australia (Turner et al. 2009) and 559 ± 6 Ma (SHRIMP) in Antarctica (Goodge et al. 2012), to 489 ± 3 Ma (\(^{39}\)Ar/\(^{39}\)Ar) in western Victoria (Miller et al. 2005) and 484 ± 8 Ma (SHRIMP) in Antarctica (Goodge et al. 2012), and perhaps even to ca 485 Ma (SHRIMP) in north Queensland (Paulick & McPhie 1999). Some microcontinental ribbons accreted directly onto the early Gondwanan margin, while others aggraded outboard. Our synthesis would suggest that VanDieland represents one of these microcontinental fragments, and was separated from the Western Lachlan Orogen by ocean crust that is now imbricated (Cayley et al. 2011). It suggests that the accretion of VanDieland must have happened after the accretion of the Dimboola Complex in western Victoria. This may have occurred during the Early Ordovician but was finalised during early Silurian Benamouin (Glen2005; Champion et al. 2008). A modern analogue is the southwest Pacific (Crawford et al. 2003b), where some microcontinental ribbons have been accreted directly onto Asia (Metcalfe 2011), while others, as in The Philippines, have come together but are yet to accrete to the Asian mainland (Yumul et al. 2003).

At the regional scale, there is a need for better paleomagnetic controls on VanDieland, but this can only come after tight age constraints can be placed on individual units. Age controls have been established in some of the glacial sequences of the Smithton Basin and associated rocks (Calver 2011; Calver et al. 2013a), but older rocks are generally less well constrained, as are supposedly equivalent sequences elsewhere. Tighter age controls have been established in some pre-570 Ma rocks on King Island and a comparison of poles with those in the Rocky Cape Zone would test the hypothesis that the two zones were from different parts of Rodinia. As well, too few Sm–Nd model ages of the basement have been determined to make statistically valid comparisons with other regions. These should be obtained not only from the Paleozoic plutonic rocks, but also from plutonic rocks on the South Tasman Rise. Only then can the hypothesis proposed here be adequately tested.

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SUPPLEMENTAL PAPER

Figure A1 Combined time–space plot (Figure 2).

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