Timing of subduction and exhumation along the Cambrian East Gondwana margin, and the formation of Paleozoic backarc basins

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ABSTRACT

The inversion of the Neoproterozoic-Cambrian passive margin of East Gondwana occurred during the early Paleozoic Delamerian-Ross orogeny. We present $^{40}Ar/^{39}Ar$ and structural data from deformed and metamorphosed Neoproterozoic clastic rocks beneath the Tasmanian ophiolite and the footwall of a high-pressure metamorphic complex in northern Tasmania. These data reveal the timing of accretionary deformation and the initiation of backarc extension along the Australian margin of Gondwana. $^{40}Ar/^{39}Ar$ analyses of muscovite from lower greenschist facies fault slices bounding the Forth metamorphic complex give plateau ages of 521.4 ± 2.5 and 520.7 ± 1.6 Ma. These data suggest that deformation within an accretionary prism off the margin of Tasmania, and possibly ocean arc collision, were under way by ca. 521 Ma. Muscovite from upper amphibolite and upper greenschist facies rocks in five locations of the Forth metamorphic complex, including retrograde shear zones, give $^{40}Ar/^{39}Ar$ cooling ages that average 508.1 ± 2.6 Ma. Identical muscovite cooling ages from rocks originally at very different metamorphic *P-T* conditions suggest rapid cooling of the Forth complex at ca. 508 Ma, due to the juxtaposition of higher-grade against lower-grade rocks. Rapid cooling is also indicated by concordant $^{40}Ar/^{39}Ar$ ages of hornblende and muscovite in the high-grade core. Cooling was probably due to rapid exhumation along extensional shear zones within a regional extensional setting that also produced the Mount Read–Mount Stavely volcanic complexes (505–495 Ma) along with rift basins in Tasmania and southeast Australia. This continental rift magmatism and extension were caused by west-dipping subduction under the Australian margin of Gondwana after the collisional phase of the Delamerian-Ross orogeny. Rollback of subduction in the Australian sector of the margin between ca. 508 and 460 Ma produced a backarc basin >1000 km wide that became the basement for the Lachlan orogen turbidites. Similar amounts of subduction rollback seem not to have occurred in Antarctica at this time (unless the record is lost), suggesting significant along-strike differences in the early Paleozoic geodynamics of the Delamerian-Ross orogenic system.

Keywords: $^{40}Ar/^{39}Ar$ dating, Gondwana, Tasmania, obduction, exhumation, Lachlan orogen, Delamerian-Ross orogen, Australia.

INTRODUCTION

The Neoproterozoic-Cambrian passive margin of East Gondwana underwent a long progression of accretionary orogenic events after the initiation of a convergent plate boundary between Gondwana and the paleo-Pacific by early Cambrian time. Inversion of the East Gondwana passive margin took place during the latter stages of the suturing of Gondwana along and within Pan-African orogenic belts, which culminated at ca. 530–510 Ma (e.g., Meert, 2003; Boger and Miller, 2004). The Cambrian passive margin in Tasmania, which was adjacent to mainland Australia and North Victoria Land, Antarctica (Fig. 1), was overthrust between 520 and 505 Ma by a mafic-ultramafic ophiolite complex along with turbidites during the early convergent stage of the Delamerian-Ross orogeny (Berry and Crawford, 1988; Elliott et al., 1993; Meffre et al., 2000). Ophiolite obduction resulted in high-pressure metamorphism of the turbidite wedge (Meffre et al., 2000; Reed et al., 2002) and was followed shortly thereafter by postcollision continental extension and magmatism at 505–495 Ma (Meffre et al., 2000; Perkins and Walsh, 1993; Black et al., 1997; Foster et al., 1998). Extension thinned and rifted the underthrust passive margin along with the ophiolite and ultimately formed an oceanic backarc basin that became the depocenter for the Lachlan orogen turbidites (Foster and Gray, 2000). The transition from obduction and shortening to oceanic backarc basin extension started a cycle of alternating convergent and divergent dynamics along the Pacific margin of Australia that formed the enigmatic Lachlan orogen by the middle Paleozoic, the New England orogen by the end of the Paleozoic, and continues to the present (Foster and Gray, 2000; Collins, 2002).

In this paper we present Ar-Ar geochronological data and summarize structural data from Neoproterozoic clastic rocks that were deformed beneath the Tasmanian ophiolite during obduction and are now exposed in the footwall of a high-pressure metamorphic complex in northern Tasmania (Figs. 2 and 3). These data bear on (1) the timing of deformation in the Delamerian-Ross orogeny, (2) the nature of exhumation of high-grade metamorphic complexes in this and other orogens, and (3) the initiation of backarc extension along the Australian margin of Gondwana. On a larger scale, these data further constrain the timing of orogenic intervals within the >20,000-km-long Paleozoic active margin of Gondwana and will help determine: (1) the possible plate tectonic settings before and during the Cambrian orogeny; (2) the age of onset of subduction, which may have varied along the margin, (3) the location of promontories and reentrants on the Neoproterozoic rifted margin,

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(4) the importance of intraorogen rifting and oceanic backarc basin formation, and (5) the nature and importance of continental blocks derived from Gondwana or other continents.

GEOLOGICAL SETTING

The western Tasmania terrane (Fig. 2) formed as an attenuated Meso- and Neoproterozoic continental block that lay along or outboard of Gondwana (Meffre et al., 2000). Mesoproterozoic continental basement in Tasmania (Berry and Burrett, 2002) does not have a direct counterpart exposed in mainland Australia or Antarctica (Elliott and Gray, 1992). This block of continental crust was probably either a promontory along the Gondwana margin or a continental ribbon rifted during the breakup of Rodinia. It records details of the inversion of the Neoproterozoic–Early Cambrian margin that are not well expressed in eastern Australia or Antarctica (Berry and Crawford, 1988).

The oldest rocks in Tasmania are quartzite and pelitic schist that contain detrital zircons with $^{207}Pb/^{206}Pb$ ages older than 1200 Ma (Black et al., 1997; Turner et al., 1998). The rocks were metamorphosed and deformed probably during the late Mesoproterozoic, based on relatively imprecise electron microprobe dates of monazite (Berry and Burrett, 2002), and intruded by shallow-level granitoids and mildly deformed at 780–760 Ma (Turner et al., 1998). Neoproterozoic continental-derived turbidite (Oonah and Burnie Formations) intercalated with basalt and intruded by syndepositional dolerite is in fault contact with the late Mesoproterozoic metasedimentary basement (Fig. 2). These mafic rocks have alkaline affinities suggesting a rift environment (Crawford and Berry, 1992) and give K-Ar biotite dates of 725 ± 25 Ma (Crook, 1979). The late Neoproterozoic successions record extension and ultimately breakup of the Australia-Antarctic margin of Gondwana (Crawford and Berry, 1992) during the fragmentation of Rodinia (Powell et al., 1994). After Neoproterozoic extension and eventual breakup, Tasmania may have been separated from Gondwana (or Laurentia) as an oceanic plateau, or may have remained attached as an attenuated promontory.

The Neoproterozoic-Cambrian continental rift to passive margin strata were highly deformed and metamorphosed in Middle Cambrian time. In Tasmania this event is known as the Tyennan orogeny, which is equivalent to the Delamerian orogeny of eastern Australia and the Ross orogeny of Antarctica. Cambrian orogeny in Tasmania and southeastern Australia was probably due to the closure of a marginal ocean basin by east-dipping subduction, which led to partial subduction of the continental margin, and west- or south-directed obduction of an ophiolite interpreted to be the roots of an oceanic island arc (Berry and Crawford, 1988; Meffre et al., 2000; Turner and Bottrill, 2001). The leading edge of the western Tasmania terrane was subducted to depths of at least 40–50 km as documented by eclogite slivers with peak metamorphic pressures of 1200–1700 MPa (Meffre et al., 2000; Turner and Bottrill, 2001). The fault-bounded eclogite slivers along with separate blueschist slivers are exposed in amphibolite facies metamorphic complexes, which themselves are fault-bounded within dominantly low-greenschist facies metamorphic sequences (Meffre et al., 2000; Turner and Bottrill, 2001). Amphibolite facies assemblages overprint most of the eclogite and blueschist facies rocks, and therefore the high *P/T* ratio assemblages are only locally preserved. Metamorphic complexes in Tasmania produced and exhumed during the Cambrian Delamerian orogeny include the Port Davey, Franklin, Mersey River, Arthur, Forth, and Badger Head complexes (Meffre et al., 2000; Reed et al., 2002) (Fig. 2). Geochemical compositions of metavolcanic and metapelitic rocks within the metamorphic complexes suggest protoliths similar to the unmetamorphosed Neoproterozoic–Lower Cambrian rift-passive margin strata in the Rocky Cape Block, northwestern Tasmania (Meffre et al., 2000; Crawford and Berry, 1992). Exposure and exhumation of the various structural levels within the metamorphic complexes appear to have occurred relatively

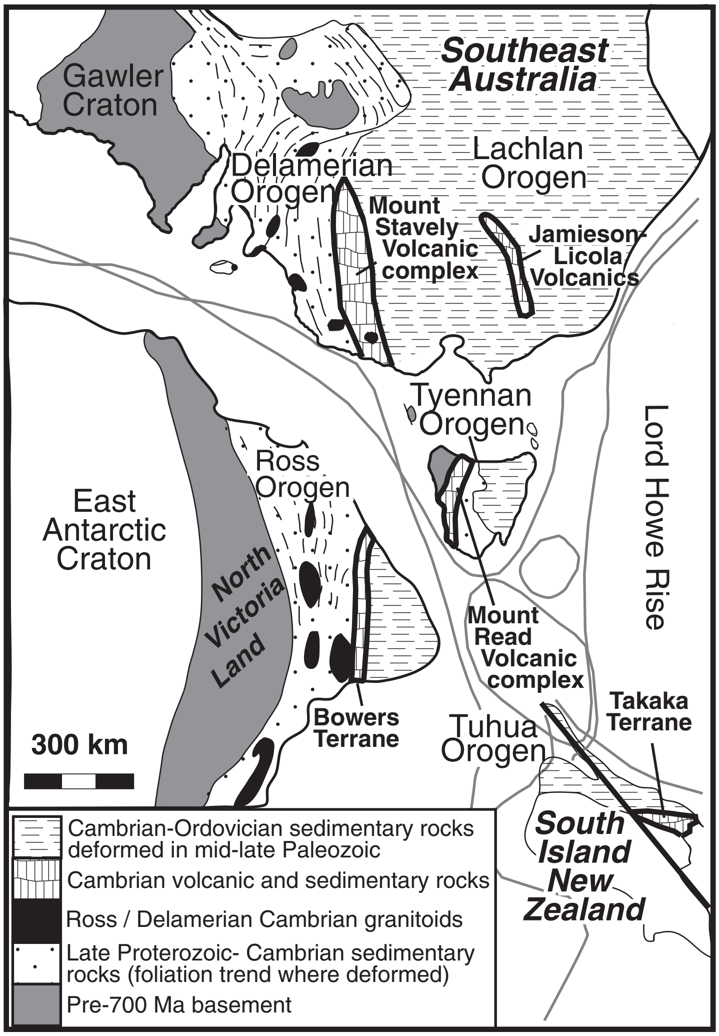

Figure 1. Reconstruction of part of the Phanerozoic East Gondwana margin—including southeast Australia, North Victoria Land, Antarctica, New Zealand, and the Lord Howe Rise—at ca. 120 Ma showing major orogenic belts and Cambrian volcanic arcs (modified from Gibson and Ireland, 1996).

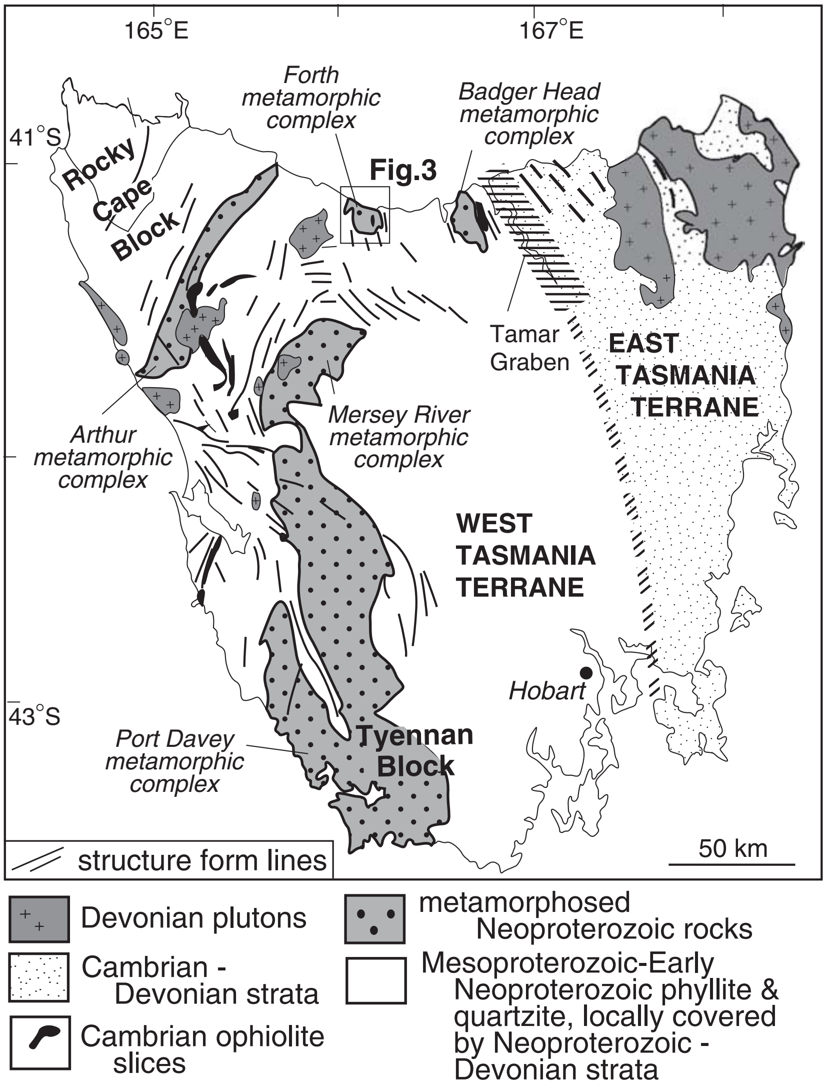

Figure 2. Geologic map of Tasmania showing Proterozoic and Paleozoic structural elements and locations of high-grade Cambrian metamorphic complexes. The box shows the location of the map in Figure 3.

rapidly based on the presence of blueschist facies minerals in Middle to Upper Cambrian clastic sediments deposited in rift basins (Turner and Bottrill, 2001).

Published U-Pb geochronology provides a framework for understanding the progression of events in the Delamerian orogeny in Tasmania. The only isotopic constraints on the age of the ophiolite are SHRIMP (sensitive high-resolution ion microprobe) U-Pb zircon analyses from late-stage tonalite within the Heazlewood River complex that give a $^{206}Pb/^{238}U$ age of 515 ± 7 Ma (Turner et al., 1998) (age referenced to standard SL13 corrected by 1% to be consistent with the updated calibration of SHRIMP dating standards by Black et al., 2003), and 513.6 ± 5.0 Ma (Black et al., 1997) (age referenced to standard QGNG and corrected by 0.4% according the Black et al., 2003, calibrations). These data provide a minimum age for the oceanic island arc that collided with Tasmania, because it is unclear when this tonalite formed in relationship to the mafic rocks (Turner et al., 1998). Eclogite metamorphism in the Franklin metamorphic complex occurred at 508.5 ± 8 Ma based on SHRIMP $^{206}Pb/^{238}U$ analyses of zircons (Turner et al., 1998) (referenced to SL13 and corrected according to Black et al., 2003). K-Ar dates of metamorphic hornblende from the Arthur metamorphic complex give an average age of 510 ± 4 Ma (Turner, 1993), which are within error of the SHRIMP zircon analyses. Metamorphic rims that grew on Precambrian zircon in garnet amphibolite from the Forth complex give a SHRIMP $^{206}Pb/^{238}U$ age of 512 ± 5 Ma (Black et al., 1997) when the data are referenced to a $^{206}Pb/^{238}U$ age of 1842 Ma for the SHRIMP standard QGNG (Black et al., 2003). Hornblende from the same garnet amphibolite gives a $^{40}Ar/^{39}Ar$ age of 508.7 ± 2 Ma (Igor Villa, 2004, written commun.; based on an age of 523.2 ± 0.9 Ma for flux monitor mmhb-1 following the calibration of Spell and McDougall, 2003; this sample was quoted as ca. 510 Ma in Turner et al., 1992). These results are consistent with previous K-Ar data from the Forth complex (McDougall and Leggo, 1965). A Cambrian age for metamorphism in the Port Davey complex is also suggested by total U-Th-Pb monazite microprobe elemental analyses with a mean age of 506 ± 25 Ma for 9 grains (Meffre et al., 2000). These data, and the overlap in errors between the youngest igneous rocks within the ophiolite and the oldest ages for peak high-pressure metamorphism, allow for subduction-obduction along the Tasmanian margin to have occurred as early as ca. 520 Ma and as late as ca. 500 Ma.

FORTH METAMORPHIC COMPLEX

The Forth metamorphic complex (Fig. 3) comprises fault-bounded slices of quartzite, amphibolite, phyllite, and schist from Proterozoic protoliths (Meffre et al., 2000). Metamorphic grade within the complex ranges from upper greenschist facies to upper amphibolite facies, with peak recorded pressures of 1300–1500 GPa and peak temperatures of 700–740 °C. Packages of rocks with similar metamorphic assemblages and *P-T* conditions are in fault contact with rocks with different metamorphic assemblages, such that the Forth complex contains rocks metamorphosed at a variety of different depths and temperatures. The Forth complex also includes an outer "rind" of highly deformed, greenschist facies metasedimentary rocks known as the Ulverstone metamorphics (Figs. 3 and 4). Highly strained, upper greenschist, quartzites, metaconglomerate, and muscovite-chlorite schist of the Ulverstone metamorphics are very well exposed along the foreshore of Bass Strait in the northwestern part of the Forth complex. The majority of the samples analyzed in this study were collected from the foreshore exposures (Figs. 4 and 5).

Forth complex metamorphic rocks (including the Ulverstone metamorphics) are in fault contact with sub-greenschist facies turbidite of the Neoproterozoic Burnie Formation and Early Cambrian Togari Group mudstone, limestone, sandstone, conglomerate, metabasalt, and chert (Fig. 4) (Berry and Gray, 2001). The polydeformed boundary between the Forth complex metamorphic rocks and Burnie Formation is a complex series of steeply dipping fault slivers that contain packages of the greenschist facies Ulverstone metamorphic rocks, Burnie Formation, and a shear zone mélange (detachment?) that includes slices of lower-grade Togari Group rocks.

The main structural break between the low-grade Burnie Formation and upper greenschist facies metamorphic rocks of the western Forth complex is the low-angle Singleton fault in the northwest exposures of the complex (Figs. 4A and 4B). This fault was originally mapped as a thrust (Burns, 1964) but carries the very low-grade (sub-greenschist facies) Burnie Formation in the hanging wall and upper greenschist facies metamorphic rocks (Ulverstone metamorphics) and platy mylonite in the footwall (Fig. 5A). Based on the approximate differences in metamorphic conditions (>200–250 °C), the rocks beneath the fault were 12–15 km deeper than those in the hanging wall if we assume a geothermal gradient of ~17 °C/km (based on the peak *P-T* metamorphic conditions). We interpret the missing metamorphic section to indicate that this fault is a normal fault or detachment zone that was ultimately responsible for late stages of exhumation of the Forth complex to shallow crustal levels after peak metamorphism. Strong north-south–trending, gently plunging (2°–20°) mineral lineations, defined by muscovite, in platy muscovite-chlorite-quartzite mylonite (Fig. 5C) and orientations of flattened and stretched clasts in metaconglomerate (Fig. 5B) indicate oblique motion on the shear zones. Zones of platy mylonite 1–10 m wide bound lenses of less-deformed, folded muscovite quartzite (Fig. 5D), contain retrograde chlorite-bearing assemblages, and increase in intensity in the direction of the higher-grade Forth complex. These relationships suggest that exhumation of the Forth complex occurred by very oblique displacement in a transtensional system, perhaps similar to the exhumation of metamorphic complexes within the Tertiary transtensional systems of the northern Cordillera (Price, 2003; Foster et al., 2003) or Walker Lane, California (Oldow, 2003).

Based on the age of metamorphism indicated by the U-Pb zircon data from the high-grade rocks (Black et al., 1997) and the presence of high-grade detritus in Middle Cambrian sedimentary basins (Turner and Bottrill, 2001), the structures described above were probably formed in Middle Cambrian time (i.e., before 500 Ma). These Middle Cambrian structures are overprinted by structures formed during several younger deformation events (Woodward et al., 1993; Elliott et al., 1993). The most obvious of these is a series of steep thrust zones (100–400 m wide) that break the northwestern Forth complex footwall and hanging-wall (Burnie Formation) boundary into a series of north-south–trending slices that produce the outcrop pattern along the foreshore (Figs. 4A and 4B). These steep faults include the Westbank and Ulverstone fault zones (Fig. 4). Slices and blocks of broken formation, metabasalt, black limestone, chert, and mudstone of the Togari Group that were caught up along the detachment are exposed in the steep fault zones. These steep fault zones may be part of a regional Late Cambrian–Early Ordovician east-west–shortening event (ca. 490–460 Ma) or related to east-directed Devonian shortening that is widespread in eastern Tasmania (Woodward et al., 1993; Elliott et al., 1993; Powell and Baillie, 1992; Reed et al., 2002). Carboniferous open folds and kinks (Powell and Baillie, 1992) and Mesozoic normal faults related to opening of the Bass Strait basins also overprint the Cambrian structures (Elliott et al., 1993; Powell and Baillie, 1992; Hall, 1998).

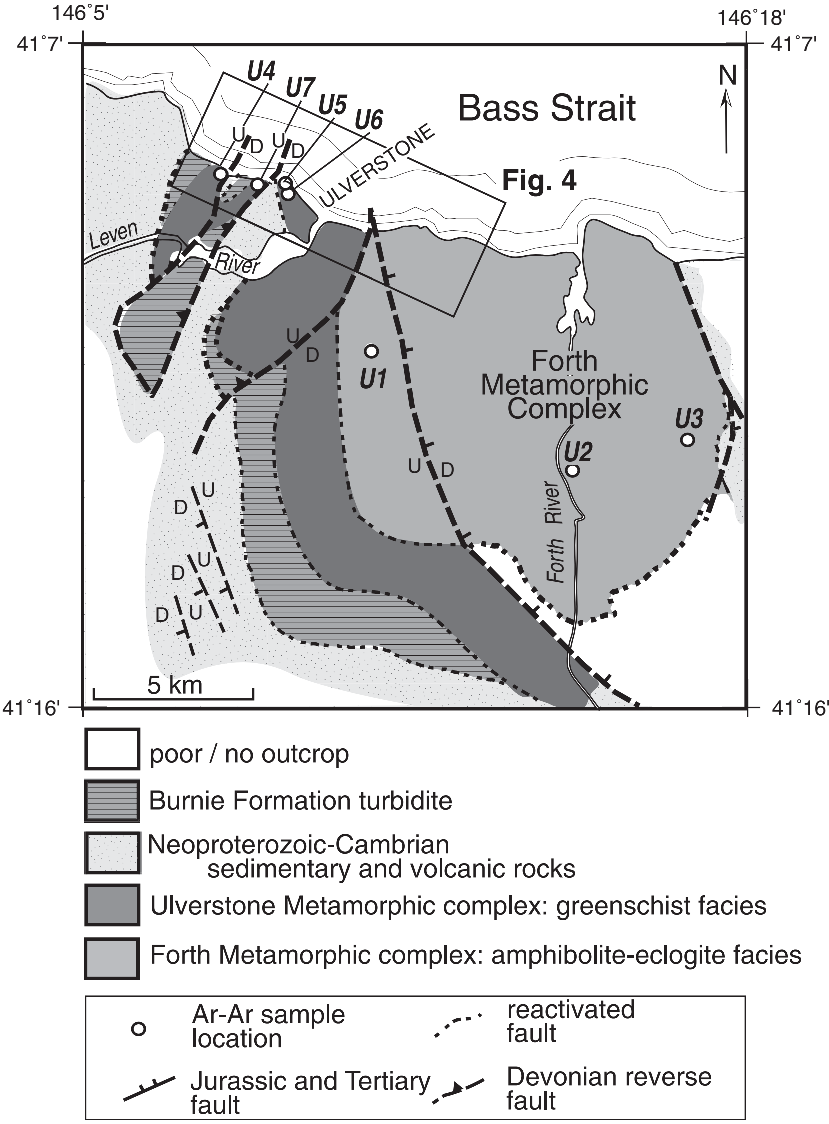

Figure 3. Simplified geological and sample location map of the Forth metamorphic complex showing the distribution of major rock units, metamorphic zones, and structures. The box shows the location of the map in Figure 4.

$^{40}Ar/^{39}Ar$ GEOCHRONOLOGY

Methods

$^{40}Ar/^{39}Ar$ analyses were performed on metamorphic mica from samples collected within the high-grade (Forth metamorphics) and lower-grade (Ulverstone metamorphics) parts of the Forth metamorphic complex, and from

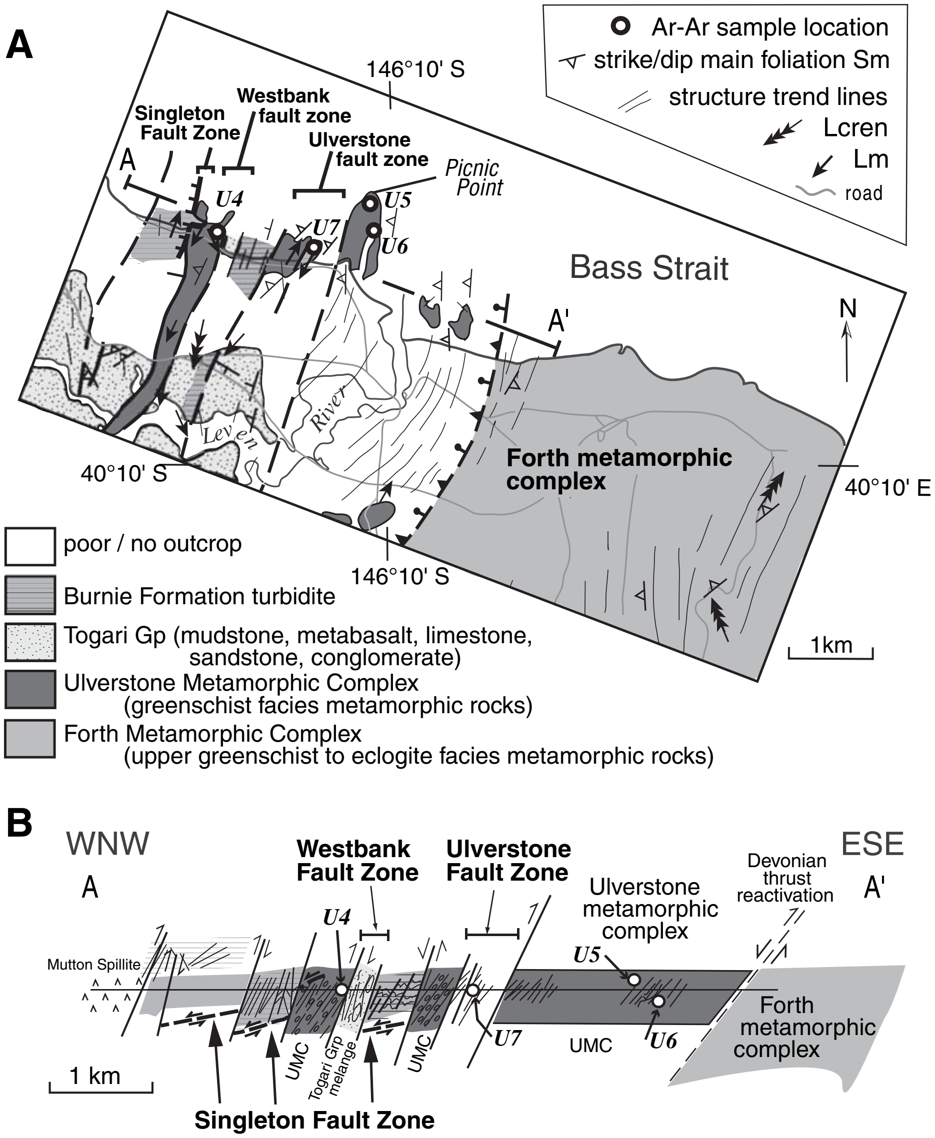

Figure 4. (A) Structural form map of the northwest part of the Forth metamorphic complex. (B) Simplified structural cross section oriented northwest to southeast (A–A′) along the shore platform. UMC—Ulverstone metamorphic complex. Sm—metamorphic layering, cleavage, schistosity, Lm—main minor fold axis attitude; Lcren—crenulation fold axis attitude/lineation.

a fault-bounded slice of Togari Group phyllite within the Ulverstone fault zone (Figs. 3 and 4).

Analyses of white mica from samples U1, U2, U5, and U7 were undertaken at La Trobe University, Australia (Foster et al., 1999), following standard methods (see McDougall and Harrison, 1999). Samples were irradiated in a core position of the IRR-1 reactor, Soreq Nuclear Research Center, Israel, for 40 h along with the flux monitor GA1550 biotite (98.5 ± 0.8 Ma: Spell and McDougall, 2003). Samples were heated in a computer-controlled double-vacuum resistance furnace with a tantalum crucible, or with a 7 W argon-ion laser. Laser step heating was done with a defocused beam and by changing the power output of the laser. Gas was expanded into a stainless steel cleanup line and purified with two 10 L per second SAES getters. Argon isotopes were measured using a VG3600 mass spectrometer with a Daly photomultiplier, operating at a sensitivity of 7×10^{-4} amp/torr. Data were corrected for machine background determined by measuring system blanks, and for mass discrimination determined by analyzing atmospheric argon. Extraction line blanks were typically $<2 \times 10^{-16}$ mole ^{40}Ar. Correction factors for interfering isotopes were determined by analyzing K_2SO_4 and CaF_2 salts irradiated with the samples, and the following values were used: $(^{40}Ar/^{39}Ar)_K = 1.56 \times 10^{-2}$, $(^{36}Ar/^{37}Ar)_{Ca} = 4.00 \times 10^{-4}$, and $(^{39}Ar/^{37}Ar)_{Ca} = 2.50 \times 10^{-2}$.

Analyses of white mica from samples U3, U4, and U6 were performed at the University of Nevada Las Vegas. Samples were wrapped in Sn foil and stacked in fused silica tubes with the neutron fluence monitor FC-2 (Fish Canyon Tuff sanidine). To be consistent with the ages calculated with GA1550, we used an age of 28.1 ± 0.4 Ma for FC-2 (Renne et al., 1998; Spell and McDougall, 2003). Samples were irradiated at the Ford reactor, University of Michigan for 6 h in the L67 position. Correction factors for interfering neutron reactions on K and Ca were determined by repeated analysis of K-glass and CaF_2 fragments included in the irradiation. Measured $(^{40}Ar/^{39}Ar)_K$ values were 1.56×10^{-2}. Ca correction factors were $(^{36}Ar/^{37}Ar)_{Ca} = 2.79 \times 10^{-4}$ and $(^{39}Ar/^{37}Ar)_{Ca} = 6.61 \times 10^{-4}$. Samples were heated using a double-vacuum resistance furnace. Reactive gases were removed by two GP-50 SAES getters prior to expansion into to an MAP 215–50 mass spectrometer. Peak intensities were measured using a Balzers electron multiplier. Mass spectrometer discrimination and sensitivity was monitored by repeated analysis of atmospheric argon aliquots from an online pipette system. The sensitivity of the mass spectrometer was 6×10^{-17} mol/mV. Line blanks averaged 17.33 mV for mass 40 and 0.06 mV for mass 36.

Results

The $^{40}Ar/^{39}Ar$ age spectra for each sample are shown in Figure 6, and a summary of the data and sample locations is given in Table 1. The complete $^{40}Ar/^{39}Ar$ isotopic data are given in Table DR1 (GSA Data Repository).[1] Samples U1 and U2 are from upper amphibolite facies garnet-muscovite-biotite-bearing schists of the core of the Forth metamorphic complex. U1 is from an outcrop containing south-dipping brittle normal faults that overprint an amphibolite facies fabric, and gives a plateau age of 507.8 ± 1.6 Ma (Fig. 6). The age spectrum for U2 is mildly discordant with apparent ages of up to ca. 550 Ma in the first ~3% of the gas released. The bulk of the steps gives apparent ages between ca. 500 and 515 Ma. The low- and high-temperature discordance between steps in U2 is probably due to the presence of minor alteration phases (e.g., chlorite, clays) observed in thin section, which resulted in some ^{39}Ar recoil redistribution. The average age

[1]GSA Data Repository item 2005023, Ar-Ar isotopic data for samples from the Forth metamorphic complex, is available on the Web at http://www.geosociety.org/pubs/ft2005.htm. Requests may also be sent to editing@geosociety.org.

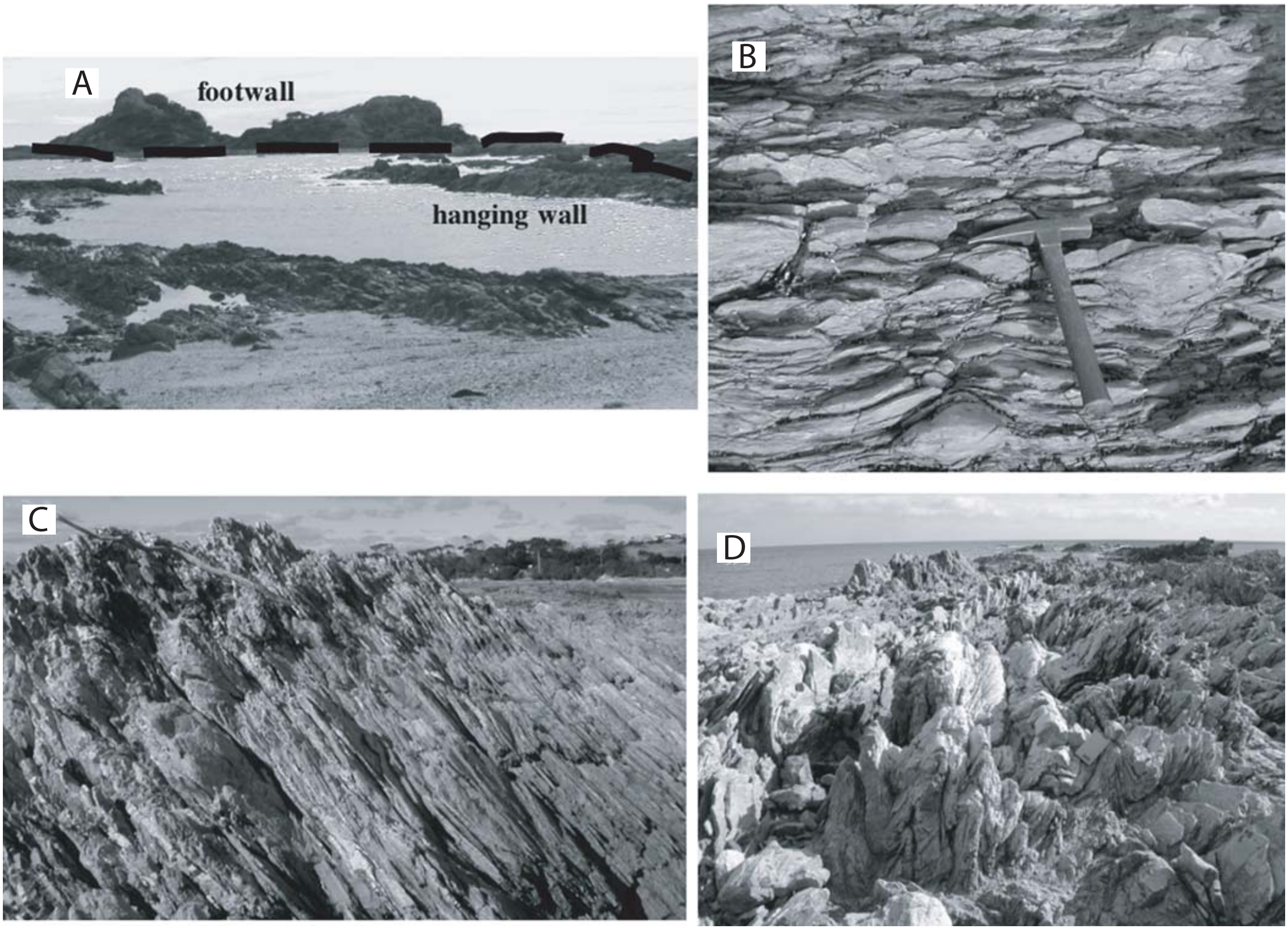

Figure 5. (A) Photograph of the low-angle outcrop pattern of the Singleton fault. (B) Photograph of strongly flattened boulders and cobbles within the greenschist facies shear zones in Ulverstone fault zone. Muscovite sample U4 was extracted from a "mica beard" that grew in a pressure shadow of one of these boulders. (C) Photograph of platy quartzite mylonite that locally overprints the S2 fabric in the northwestern Forth complex. Sample U6 is from one of these shear zones. (D) Photograph of folded muscovite quartzite at Picnic Point showing multiple fold interference within a lens of less-deformed greenschist facies rocks within the shear zones that bound the Forth complex. Sample U5 is muscovite separated from the quartzite in this location.

of steps between 7% and 100% gas released in U2 is 508.8 ± 2.4 Ma, which is interpreted as a maximum cooling age for this white mica. U3 is a chlorite-muscovite-quartzite from an upper greenschist facies slice of the Forth complex. The plateau age, which represents ~60% of the ^{39}Ar released, of muscovite from this sample is 521.4 ± 2.5 Ma. The low-temperature steps in U3 are older than the plateau age and may be due to excess argon incorporated into minor chlorite intergrowths within the muscovite, or recoil redistribution into the chlorite. Samples U4, U5, and U6 are from upper greenschist facies quartz cobble conglomerates and muscovite quartzite that are highly strained, exhibiting strong flattening fabrics and north-south–stretching lineations (S2 and L2, respectively: Berry and Gray, 2001). U4 and U5 are from less-deformed lenses in the shear zone. U4 consists of muscovite from a pressure shadow around a flattened cobble, and U5 is from a folded quartzite. U6 is from a fine-grained platy quartzite mylonite, with a greenschist facies fabric that overprints the S2 fabric. Samples U4, U5, and U6 yielded plateau ages of 508.7 ± 2.5 Ma, 507.7 ± 1.6 Ma, and 508.6 ± 2.5 Ma, respectively. Sample U7, from a highly strained lowest greenschist facies Togari Group phyllite within mélange in the Ulverstone fault zone (Fig. 4), gives a plateau age of 520.7 ± 1.6 Ma.

DISCUSSION

Deformation and Exhumation History

The argon closure temperature for metamorphic white mica varies with composition but ranges from ~450 to 350 °C for cooling rates of 50–5 °C/m.y. for grains with average diffusion dimensions of ~5.9 microns (Lister and Baldwin, 1996; McDougall and Harrison, 1999), which is larger than the mica grain size from any sample in this study. $^{40}Ar/^{39}Ar$ analyses of muscovite from the amphibolite and upper greenschist facies rocks within the Forth complex, therefore, yield postmetamorphic cooling ages. Lower peak metamorphic temperatures (330–350 °C) for the Togari Group phyllite from within the Ulverstone fault zone (U7) suggest that the metamorphic micas in this sample grew at or below their closure temperature. This assumes that the closure temperature for these phengitic white micas is at least 350–380 °C as summarized above. Therefore, the plateau age of the mica in sample U7 (520.7 ± 1.6 Ma) most likely

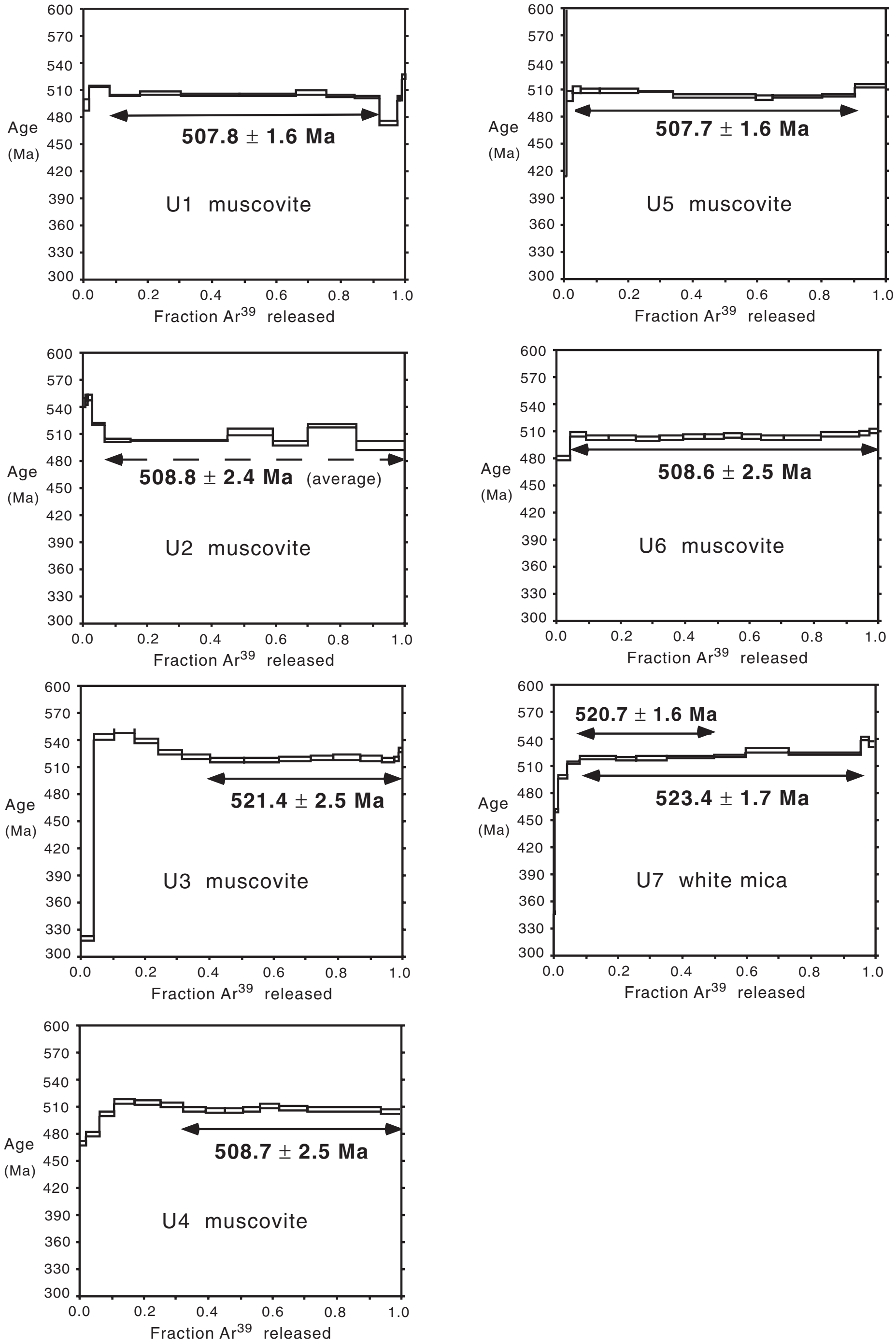

Figure 6. $^{40}Ar/^{39}Ar$ age spectra for samples from the Forth metamorphic complex.

reflects the time of prograde metamorphism for the Togari Group in this part of Tasmania. Muscovite within the lower-grade fault slice on the west side of the Forth complex (U3) may also be giving an age approximating the time of prograde metamorphism (521.4 ± 2.5 Ma), because metamorphic assemblages in this sample suggest a temperature of 350–450 °C. It is also possible that the older age from the phyllite (U7) is due to minor residual detrital mica that was not recrystallized during deformation. A small fraction of residual radiogenic ^{40}Ar in detrital grains could explain the gentle increase in age from ca. 521 Ma to older than 530 Ma in the latter part of the age spectrum, although this gradient could indicate that metamorphism started closer to ca. 530 Ma. No detrital mica was observed in thin section for sample U7, due to the very strong degree of deformation and recrystallization. Moreover, this sample gives an age equal to that of sample U3, which is from higher-grade rocks. We argue, therefore, that the apparent age of U7 yields information about metamorphism and deformation.

Ar-Ar results from the lower-grade samples (U7 and U3), therefore, indicate that thrusting and metamorphism of the western Tasmania continental margin was under way by 520.9 ± 3.6 Ma, which is the 2σ weighted average of the two plateau ages, including errors in the ^{40}K decay constant (Renne et al., 1998). This age is within error of the U-Pb zircon dates from the high-grade parts of the Forth and Franklin metamorphic complexes (Black et al., 1997), when the zircon data are corrected for revisions to the SHRIMP dating standards (Black et al., 2003), and possible external variations in the ^{40}K decay constant are taken into consideration (Min et al., 2000). Postcollision extension and exhumation was under way by 508.1 ± 2.6 Ma, based on the weighted average (2σ including decay constant error) of the mica cooling ages of U1, U2, U4, U5, and U6. Exhumation of the high-pressure metamorphic rocks by Middle Cambrian time is consistent with the presence of metamorphic detritus, including blue amphiboles, in the rift basins of this age in western Tasmania (Turner and Bottrill, 2001).

Using the approximate closure temperatures of the isotopic systems for the U-Pb and $^{40}Ar/^{39}Ar$ data, we are able to place some boundaries on the rate of cooling for the northern Forth metamorphic complex. Upper amphibolite facies metamorphism in the high-grade parts of the Forth complex took place at 511 ± 4 Ma, based on the U-Pb zircon age, which records peak metamorphic temperatures of 700–740 °C (Meffre et al., 2000). High-grade rocks in the northern Forth complex cooled through Ar closure for hornblende (~530 °C for a cooling rate of 50 °C/m.y.: McDougall and Harrison, 1999) at 509 ± 2 Ma, based on the $^{40}Ar/^{39}Ar$ hornblende age (I. Villa, 2004, written commun.; Turner et al., 1992) and muscovite argon closure (400–350 °C) by 508.1 ± 2.6 Ma. Directly comparing these three mineral ages and closure temperatures is only possible with the $^{40}Ar/^{39}Ar$ hornblende and muscovite ages, because of potential decay constant errors introduced when comparing the $^{40}Ar/^{39}Ar$ and U-Pb data. The $^{40}Ar/^{39}Ar$ hornblende and muscovite ages are indistinguishable within error and, given that they record cooling over a temperature interval of 100–180 °C, indicate very rapid cooling. Very rapid rates of cooling resulting in concordant hornblende and muscovite $^{40}Ar/^{39}Ar$ ages are common for metamorphic core complexes that experienced tectonic denudation (Foster

TABLE 1. SAMPLE DETAILS AND SUMMARY OF AR-AR DATA FROM THE FORTH METAMORPHIC COMPLEX

Sample number	Location	Rock unit/mineral assemblage	Mineral	Ar-Ar age ± 1 σ (Ma)
U1 (00-04)	Forth metamorphic complex footwall, B15 road cut: 41°12.386′S, 148°11.049′E	Garnet, muscovite, quartz schist	Muscovite (>150 µm grains)	507.8 ± 1.6
U2 (88-490)	Forth metamorphic complex footwall, Sayers Hill Road	Garnet, muscovite, quartz	Muscovite (>150 µm grains)	508.8 ± 2.4
U3 (01-07)	Forth metamorphic complex footwall, Lower Barrington Road: 41°13.088′S, 148°17.132′E	Chlorite, muscovite, quartzite	Muscovite (>150 µm grains)	521.4 ± 2.5
U4 (01-01)	Westbank fault zone, Ulverstone foreshore: 41°08.511′S, 146°08.508′E	Ulverstone metamorphics, conglomerate with flattened pebbles; muscovite quartzite	Muscovite (>150 µm grains)	508.7 ± 2.5
U5 (P2)	Picnic Point, Ulverstone foreshore: 41°08′S, 146°09′E	Chlorite, muscovite, quartzite	Muscovite (>150 µm grains)	507.7 ± 1.6
U6 (01-12)	Picnic Point, Ulverstone foreshore: 41°08.719′S, 146°09.771′E	Chlorite, muscovite, quartz schist parting from platy mylonite	Muscovite (>150 µm grains)	508.6 ± 2.5
U7 (88-430)	Ulverstone fault zone, Ulverstone foreshore:	Togari Group: muscovite, chlorite, quartz phyllite	Muscovite concentrate (>20–50 µm grains, with <10% quartz)	520.7 ± 1.6

Note: Error in age reflects uncertainty in J curve and flux monitor age but not error in decay constant (5.444×10^{-10} yr^{-1}).

and John, 1999) and could also be consistent with rapidly eroding transpressional orogens undergoing tectonic extrusion (Batt et al., 2000). The average peak metamorphic temperature (700–740 °C) and depth (40–45 km) for the Forth complex garnet amphibolite (Meffre et al., 2000) suggests a metamorphic geothermal gradient of 17 ± 2 °C/km. Therefore, the rapid cooling rates indicated by the $^{40}Ar/^{39}Ar$ data must correspond to exhumation rates of greater than 1 km/m.y.

The cause of rapid exhumation is debated for most high-pressure terranes, and the relatively poor outcrop in northern Tasmania means that the structural data in the area are incomplete. However, rapid cooling and exhumation were most likely the result of extension and metamorphic core complex formation (Hall, 1998), driven by buoyant extrusion, transtension, or a major change in plate dynamics. Rapid cooling rates implied by the coincidence of mica ages from different structural levels across the Forth complex and the bounding shear zones (Ulverstone metamorphics) suggest tectonic exhumation was ongoing at ca. 508 Ma. The final exhumation of the Forth complex probably occurred on the low-angle Singleton detachment. Tectonic unroofing of the Forth complex is also consistent with the tectonic setting that formed large rift basins in Middle Cambrian time, which contain detritus from this and other Tasmanian metamorphic complexes (Turner and Bottrill, 2001). Continued postcollisional extension led to the formation of the Mount Read volcanic complex at 505–495 Ma, which was most likely generated in a continental backarc rift setting by decompression melting of the lithospheric mantle (Münker and Crawford, 2000).

The $^{40}Ar/^{39}Ar$ results, therefore, suggest that shortening and metamorphism of the western Tasmania continental margin began by 520.9 ± 3.6 Ma and was complete by 508.1 ± 2.6 Ma. This gives a span in time of ~7–19 million years for initial tectonic burial, arc collision, partial subduction of the West Tasmania Block to ≤40–50 km during ophiolite obduction, and exhumation of high-grade rocks including amphibolites, blueschists, and eclogites (Fig. 7).

Tectonic Implications

These constraints on deformation and exhumation have implications for the Cambrian tectonic evolution of the Australia-Antarctic segment of the active margin of Gondwana, which extended from Australia to South America (e.g., Rapela et al., 1998). It is not clear when subduction along the paleo-pacific margin of Gondwana started, but it was likely after ca. 560–550 Ma (e.g., Veevers, 2003; Goodge, 1997; Rapela et al., 1998). This change from passive-rift margin to subduction occurred during the final amalgamation of Gondwana, which was complete by ca. 535–520 Ma (Meert, 2003; Jacobs et al., 1998; Boger and Miller, 2004). Within central and northern Australia intraplate transpressive deformation in the Petermann Ranges orogeny and King Leopold Ranges took place at ca. 550–530 Ma (Veevers, 2000; Camacho et al., 2002). This intraplate transpression was probably due to regional stresses within East Gondwana associated with the later stages of Mozambique Belt–Kuunga collisions and therefore was not directly related to subduction along the East Gondwana margin.

The convergent stage of the Delamerian orogeny on mainland southern Australia started at or before 514 ± 5 Ma, based on U-Pb zircon geochronology from syntectonic granitoids (Rathgen gneiss in South Australia) (Foden et al., 1999). Oblique convergence and transpressional deformation in the Ross orogen, Antarctica, started prior to ca. 530 Ma (Goodge et al., 1993a; Goodge, 1997), and evidence from magmatic rocks suggests that west-dipping oblique subduction initiated under East Antarctica by ca. 550–540 Ma (Goodge et al., 1993b; Goodge and Dallmeyer, 1996; Encarnación and Grunow, 1996; Flöttmann et al., 1998; Veevers, 2000). The onset of the main orogenic phase of the Ross orogen, however, probably postdates 530–520 Ma (Encarnación and Grunow, 1996). Therefore, the present evidence for the onset of significant Cambrian convergent deformation is 530–520 Ma in Antarctica, 520.9 ± 3.6 Ma in Tasmania, and older than 514 ± 5 Ma on the southeast Australia mainland. Rapid opening and deposition within the Kanmantoo Rift and basin by ca. 526 Ma predates the convergent stage of the Delamerian orogeny in South Australia (Jago et al., 2003). The Kanmantoo Rift is filled with >8 km of turbidite sourced mainly from the Pan-African-aged (600–500 Ma) and Grenville-aged (1200–1000 Ma) orogenic belts, based on U-Pb ages of detrital zircons (Ireland et al., 1998). The Kanmantoo Rift and adjacent ocean(?) basin to the east most likely opened as a result of oblique convergence between Gondwana and the paleo-Pacific plates, which started possibly as early as 550–530 Ma (Veevers, 2000). The opening of this basin may indicate that subduction far off the Australian margin was taking place as early as 550–530 Ma (Fig. 8A). The variations in style and timing of initial tectonism associated with

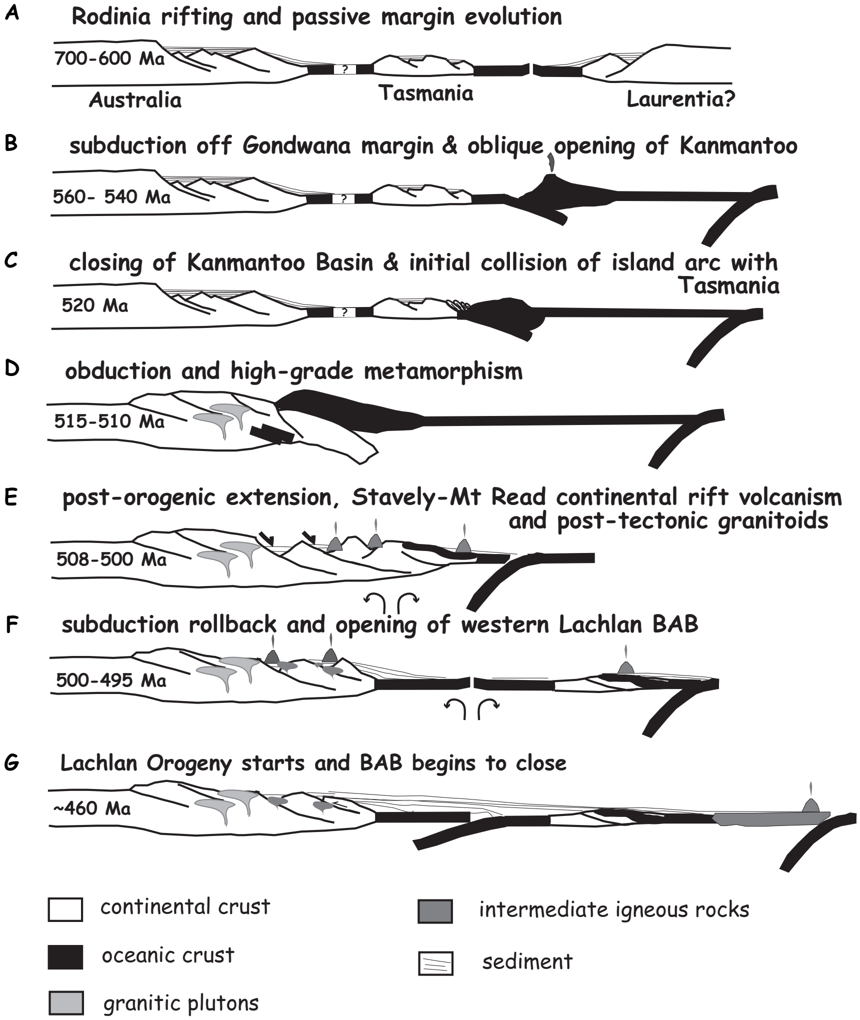

Figure 7. Cross-sectional model for the interaction between mainland Gondwana, Tasmania, and oceanic elements during the Neoproterozoic through Silurian time interval. (A) Rifting of Rodinia with Tasmania being stranded as a microcontinent off the margin of Gondwana. (B) Initiation of oblique subduction in the paleo-Pacific ocean off the margin of Gondwana. (C) Start of the collision of Tasmania with an oceanic island arc, probably on an eastward-dipping subduction zone (Crawford and Berry, 1992). (D) Obduction of the Tasmanian ophiolite and peak metamorphism (Meffre et al., 2000). (E) Rapid extensional failure of the Delamerian-Ross collision zone, exhumation of high-grade metamorphic complexes, continental rift volcanism, and west-dipping subduction under Australia. (F) Rollback of subduction causing further extension, continental rift and arc volcanism, and rifting of the Gondwana margin, forming the Lachlan backarc basin. (G) Shortening of the Lachlan backarc basin and formation of the western Lachlan orogen accretionary prism.

subduction along the paleo-Pacific margin of Gondwana could be explained by a strongly obliquely convergent margin setting where the accommodation mechanisms (e.g., transtension, transpression, continental or oceanic volcanism, etc.) were variable along the margin depending on the distance from the subduction zone, dip of subduction, etc. (e.g., Foster et al., 1999; Goodge and Dallmeyer, 1996; Veevers, 2003).

The variable Early Cambrian histories along the Australian-Antarctic margin (pre–508 Ma) may indicate that a transform boundary existed between and linked separate subduction zones, as suggested by some authors (e.g., Münker and Crawford, 2000; Flöttmann et al., 1998). Even the subduction polarity may have been different prior to the ophiolite obduction event. Subduction was probably west-dipping under Antarctica, based on the locations of the Granite Harbor intrusives in North Victoria Land (e.g., Ricci et al., 1997), and east-dipping away from Tasmania and Australia, allowing for southwest-directed ophiolite obduction (Fig. 8). However, opposite-dipping paleo-Pacific subduction zones are not required if the Tasmania–island arc ophiolite collision occurred in a single, but complex, suprasubduction setting, as is common in the southwestern Pacific today (Fig. 8B). In this case, differences between Tasmania, Australia, and Antarctica were caused by the variable splitting and subduction of a backarc basin, which grew when the Kanmantoo Rift formed.

Extension and rapid exhumation of the Forth complex by ca. 508 Ma was followed in Tasmania by the development of the Mount Read, calc-alkaline, volcanic-plutonic complex at ca. 505–495 Ma (Perkins and Walsh, 1993; Black et al., 1997). Similar-aged extensional events and calc-alkaline volcanic rocks are exposed on mainland Australia in the Mount Stavely volcanic belt (Figs. 1 and 9), as well as within the Bowers Rift in Antarctica (Münker and Crawford, 2000). The geochemical signatures of the Mount Read and Mount Stavely volcanics indicate that they are continental rift magmas that formed in an extensional setting after the main shortening phase of the orogeny (Crawford and Berry, 1992; Münker and Crawford, 2000; Foden et al., 2002). These volcanic provinces are therefore not directly "Andean" continental margin arc volcanics as some previous models have suggested. Another Middle Cambrian andesitic oceanic volcanic arc, with suprasubduction zone affinities and Cambrian sedimentary rocks different from those within the Mount Stavely and Mount Read volcanics, existed to the east of the western Lachlan basin (Fig. 9). This arc is exposed in the Jamieson-Licola inliers north of the present position of Tasmania (Cayley et al., 2002; Spaggiari et al., 2002, 2003a, 2003b). The Mount Stavely volcanics and Jamieson-Licola volcanics are now separated by deformed turbidites of the western Lachlan orogen (Foster and Gray, 2000), which were deposited mainly on oceanic backarc basin crust that is dominantly 505–495 Ma, based on dates of the basalts and faunal assemblages in overlying sediments (Foster and Gray, 2000; Spaggiari et al., 2003a; Crawford and Keays, 1987; VandenBerg et al., 2000). The western Lachlan basement also includes fragments of slightly older Middle Cambrian oceanic crust along with boninitic crust of forearc affinity (Crawford and Cameron, 1985; Crawford and Keays, 1987; Spaggiari et al., 2003a).

Formation of the 505–495 Ma Mount Read–Mount Stavely backarc, rift magmatic province was most likely related to extension associated with rollback or steepening of west-dipping subduction under the Australian part of the Gondwana margin after the shortening stage of the Delamerian orogeny (Figs. 7 and 8). Continental extension above and behind the new subduction zone would give rise to the geochemical and isotopic characteristics of this arc, the opening of rift basins, and associated exhumation of the high-grade core of the Delamerian mountain belt by extension (e.g., Tasmanian metamorphic complexes). The ca. 500 Ma oceanic basin that makes up the western and central Lachlan orogen is interpreted to be the result of backarc spreading and rollback of a west-dipping subduction zone from ca. 505 Ma to 495 Ma (Foster et al., 1999; Foster and Gray, 2000). Backarc extension may have split the Jamieson-Licola volcanics away from the Mount Stavely–Mount Read province. This would be consistent with hypotheses suggesting that a northern extension of Tasmanian Proterozoic basement underlies the easternmost part of the western Lachlan orogen (Cayley et al., 2002). However, the fact that the two magmatic provinces were active over the same time interval on opposite sides of a forming oceanic basin suggests that they were not originally connected (Spaggiari et al., 2003a, 2003b). This implies a complex arrangement of oceanic and continental microplates developed off the Delamerian-Ross margin of Gondwana starting at ca. 505 Ma and continuing to ca. 490 Ma (Figs. 8 and 9).

The change from proposed east-dipping subduction away from the Australian margin to west-dipping subduction under the Australian margin after ophiolite obduction at ca. 521–510 Ma, therefore, started a cycle of oceanic backarc basin formation, and closure by subduction, that continued throughout Phanerozoic time (Foster and Gray, 2000; Collins, 2002). Subduction rollback at 508–495 Ma gave rise to postorogenic extension of the Delamerian orogen, exhumed the high-grade metamorphic complexes of western Tasmania, initiated the Gondwana margin calc-alkaline and alkaline magmatic provinces and island arcs, and produced the western Lachlan oceanic backarc basin. It remains unclear if part of the continental crust of the West Tasmania Block was partly separated from Gondwana during the rollback stage (Fig. 7). Delamerian basement in New Zealand (Gibson and Ireland, 1996) could have also been separated from Gondwana at this time.

A second east-west–shortening event followed the Mount Read magmatic and extensional events in Tasmania in Late Cambrian time (ca. 495–490 Ma) (Crawford and Berry, 1992), and a similar-aged event affected the Delamerian in southeastern mainland Australia (VandenBerg et al., 2000). This could have been caused by a regional change in plate motions that also ended backarc basin spreading in the western Lachlan. No mafic basement has been identified in the Lachlan that is younger than ca. 495 Ma. Also at ca. 495 Ma, plutonism in the Delamerian and Ross orogens changed to extension-related A-type compositions (Borg and De Paolo, 1991; Foden et al., 1990; Allibone et al., 1993).

Late Ordovician–Silurian tectonism began to close the western Lachlan backarc basin (Fig. 7), resulting in an average of ~70% shortening in the thick Cambro-Ordovician turbidite sequence on the basaltic crust between 460 and 420 Ma (Foster et al., 1999). Recent identification of Silurian structures in deformed Ordovician turbidites in northeastern Tasmania is consistent with this model (Reed, 2001; Reed et al., 2002). Tasmania was, therefore, an integral part of two major orogenic intervals along the Gondwana

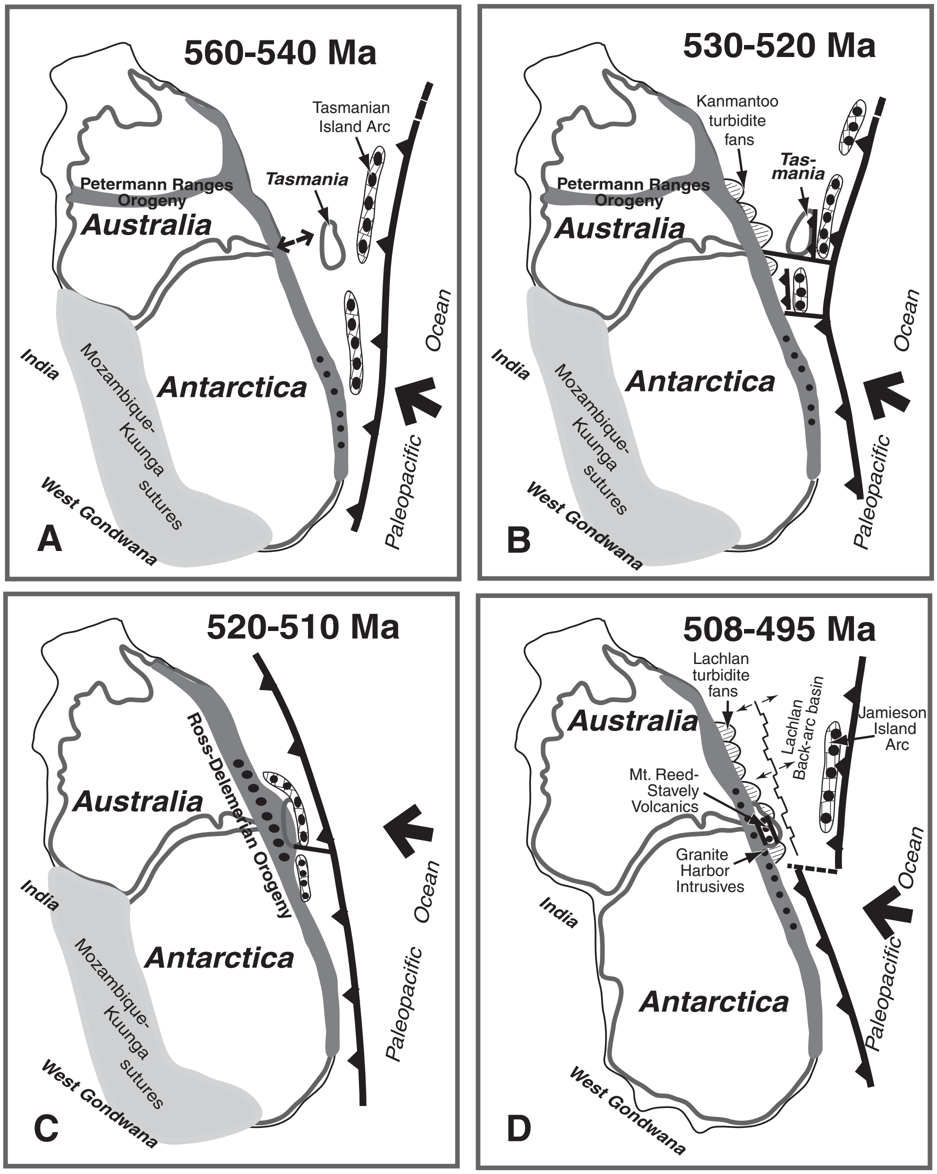

Figure 8. Evolution of the Australia-Antarctic part of the Gondwana margin during Cambrian time. (A) Oblique convergence and subduction of the paleo-Pacific starts the inversion of the passive margin, magmatism in the southern and central Transantarctic Mountains, and oblique opening of the Kanmantoo Rift in Australia. (B) Growth of island arcs and closure of backarc basin by subduction and initial deformation in Tasmania. (C) Complete obduction of the Tasmanian ophiolite, shift to near orthogonal convergence, and major shortening in the Delamerian-Ross orogen. (D) Postorogenic extension and backarc basin formation off the Australian margin above a steep west-dipping subduction zone. Shaded area is the Delamerian-Ross orogen and other contemporaneous orogenic belts. Black circles indicate areas of active magmatism; circles in hatched areas represent oceanic volcanic arcs.

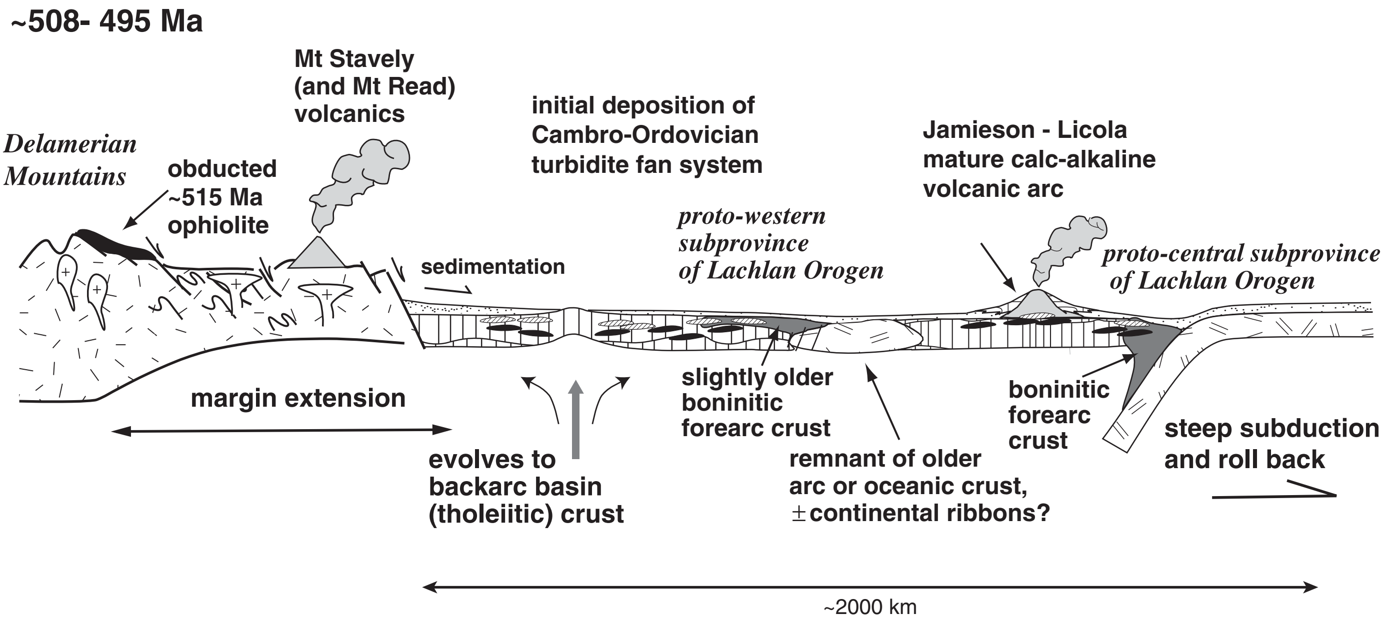

Figure 9. Cross section illustrating the major tectonic elements associated with the growth of the Lachlan backarc basin along the Australian part of the Gondwana margin at 508–495 Ma (modified from Foster and Gray, 2000).

margin: the Delamerian-Ross orogeny, when Tasmania collided with an oceanic island arc that was thrust onto the Gondwana passive margin; and the Lachlan orogeny (460–380 Ma), when extended parts of the Delamerian orogen in Tasmania were forced back onto Australia-Antarctica during closing of a backarc basin that reactivated Delamerian-Ross structures.

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