An Archaean province in the southern Prince Charles Mountains, East Antarctica: U–Pb zircon evidence for c. 3170 Ma granite plutonism and c. 2780 Ma partial melting and orogenesis

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Abstract

U–Pb ion microprobe analysis of zircon indicates that the oldest rocks from the southern Prince Charles Mountains, east Antarctica, consist of granitoid rocks of the Mawson Suite that were emplaced between 3185 and 3155 Ma. These granitoids are tectonically intercalated with the Menzies Series, a thick sequence of Cr-quartzite, marble, metaconglomerate and kyanite-bearing metapelite. Together these rocks were deformed during a single Late Archaean tectonothermal event that occurred between 2790 and 2770 Ma based on the ages of two syn-tectonic leucosomes. This event deformed the terrane via south to southwest-directed shortening and metamorphosed the rocks to upper-amphibolite facies. A sample of post-tectonic pegmatite with an age of 2645 Ma further confirms the Archaean age of deformation and metamorphism recorded by these rocks. These new age constraints demonstrate that: (1) the sialic basement (Mawson Suite + Menzies Series) of the southern Prince Charles Mountains formed entirely in the Archaean and, (2) these rocks do not share a common geologic history with the other known exposures of pre-Gondwana basement from the East Antarctic Craton, nor with the rocks of the previously continuous Gawler Craton of southern Australia. This latter point implies that the East Antarctic Craton probably did not exist as a single entity in the Archaean and Early Proterozoic, but rather formed in the Middle to Late Proterozoic via the amalgamation of a number of discrete lithospheric blocks.

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1. Introduction

East Antarctica as it appears today consists of an ancient cratonised core surrounded on both the Dronning Maud Land and Transantarctic margins by a collage of younger orogenic belts and accreted terranes (Fig. 1). On the Dronning Maud Land side these addi-

tions occurred during the collision and suturing of Gondwana and included the accretion and variable overprinting of the Mesoproterozoic Maud and Archaean Grunehogna Complexes (e.g. Jacobs *et al.*, 2003). Further to the east, the Archaean Napier and Proterozoic Fisher and Rayner Complexes of Enderby and MacRobertson Land, together with the Archaean Vestfold Hills Complex of Princes Elizabeth Land, are also interpreted to have been accreted into their present positions as a result of Gondwana tectonics (e.g. Boger and Miller, 2004). On the opposite Transantarctic margin of Antarctica, the outward growth of the continent reflected the formation of an active and long-lived subduction margin

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Fig. 1. (a) Map of present day Antarctica showing the distribution of outcrop and the location of study area (boxed) with respect to the major geographic features, named regions and permanent Australian stations. Enlargement of the study area is illustrated in Fig. 2. (b) Pre-rift fit of Australia and Antarctica (c. 70 Ma) showing the extent of the East Antarctic Craton and the distribution of accreted terranes. N, R, T, W refer to the regions of exposed rocks that formed part of the pre-Gondwana East Antarctic Craton.

subsequent to Gondwana amalgamation (Cawood, 2005).

In contrast to these areas, the interior of East Antarctica is thought to have formed a single intact craton for at least the last 1000 million years of Earth's history. The rocks from East Antarctica's cratonic interior (Fig. 1) crop out in the hinterland of the Transantarctic Mountains (Goodge and Fanning, 1999; Goodge et al., 2001), form isolated rock exposures along the Wilkes Land and Terre Adélie coast (Peucat et al., 2002), and define some of the southernmost rock exposures in the southern Prince Charles Mountains (Boger et al., 2001; Mikhalsky et al., 2001). Prior to the Eocene separation of Australia from Antarctica, this region was continuous with the Gawler Craton of South Australia (Flöttmann and Oliver, 1994; Oliver and Fanning, 1997). Collectively this larger craton is known as the Mawson Continent (Fanning et al., 1996) and is thought to have formed one of the Earth's oldest and largest cratonic elements (Fig. 1).

In this contribution, we describe the Archaean intrusive and deformation history of the Ruker Province of the southern Prince Charles Mountains. This region is thought to form part of the cratonised core of East Antarctica and is the only known exposure of such rocks between longitudes 30°W and 100°E (Fig. 1b). Russian geologists visited the Ruker Province for the first time in the late 1950s. The region was again visited by Australian geologists in the 1960s and early 1970s. The findings of these expeditions were partly published in the international literature in the 1960s (e.g. Trail, 1963a), although most results appeared later in a series of conference volumes and Australian Bureau of Mineral Resources reports (Ruker, 1963; Trail, 1963; Tingey and England, 1973; England and Langworthy, 1975; Tingey et al., 1981; Grew, 1982; Hofmann, 1982; Kamenev, 1982; Lopatin and Semenov, 1982; Tingey, 1982). This work, recently reviewed by Tingey (1991), Kamenev et al. (1993) and Mikhalsky et al. (2001) provided a broad understanding of the distribution, grade, and geochemistry of the observed rocks, but lacked the robust geochronological data required to temporally constrain the observed intrusion and deformation events. We build on these studies by presenting a description of the major deformation events recognised along the southern Mawson Escarpment. We couple these observational results with Sensitive High Resolution Ion-Microprobe (SHRIMP) U–Pb zircon data acquired from some of the major Archaean felsic intrusions as well as from leucosome and pegmatite bodies that form various cross-cutting relationships with the observed structures. These data temporally constrain the major igneous and orogenic events observed in this part of East Antarctica and

provide new constraints on the make-up and evolution of the Mawson Continent.

2. East Antarctic Basement Provinces

The modern Antarctic icesheet covers approximately 95% of East Antarctica. As a consequence exposures of pre-Gondwana basement rocks are limited to four main areas located mostly on the periphery of the East Antarctica (Fig. 1). These regions are: (1) the Nimrod Province; (2) the Terre Adélie Province; (3) the Wilkes Province and; (4) the Ruker Province. All are widely separated and often geologically distinct regions of Archaean and Palaeo- to Mesoproterozoic crust (Fig. 1). They crop out as exposed basement in the central Transantarctic Mountains (Nimrod Province), as inland ranges and nunataks in the southern Prince Charles Mountains (Ruker Province) and as coastal exposures between 100°E and 150°E (Wilkes and Terre Adélie Provinces).

Of the four basement provinces, rocks from the Nimrod and the Terre Adélie Provinces share the most in common. In both regions, Palaeoproterozoic orogenesis produced a high-grade structural and metamorphic overprint dated between 1730 and 1690 Ma. In the Nimrod Province, Palaeoproterozoic metamorphism and syn-tectonic magmatism (1730–1720 Ma; Goodge et al., 2001), reworked Archaean gneisses initially emplaced at c. 2980 Ma and which were also metamorphosed during an Archaean high-grade event dated at c. 2955 Ma (Goodge and Fanning, 1999). The rocks from the Terra Adélie Province (Fig. 1b) were deformed and metamorphosed slightly later—between 1710 and 1690 Ma (Peucat et al., 1999; Oliver and Fanning, 2002). Although this event is temporally similar to that described from the Nimrod region, the crustal depth of orogenesis was considerably shallower. Low-pressure metamorphism is implied for the Terra Adélie Province (Stüwe and Oliver, 1989) as compared to the middle- to lower-crustal depths inferred for the Nimrod Orogeny (Goodge et al., 2001). Palaeoproterozoic orogenesis along the Terre Adélie Coast variably overprinted both Palaeoproterozoic sediments (deposited after 1760 Ma) and Late Archaean to Early Palaeoproterozoic basement (2500–2400 Ma), both of which have direct correlates with rocks from the Gawler Craton of South Australia (Fanning et al., 1988; Flöttmann and Oliver, 1994; Oliver and Fanning, 1997, 2002; Peucat et al., 1999; Ménot et al., 2005). The Archaean and Palaeoproterozoic sequences of the Gawler craton were deformed and metamorphosed between 1730 and 1700 Ma (Daly et al., 1998; Hoek and Schaefer, 1998; Vassallo and Wilson, 2001, 2002), an interval that overlaps temporally with the

Palaeoproterozoic events recognised in the Terra Adélie Province as well as the more distal Nimrod Province. The correlation of rocks between southern Australia and the Terre Adélie region is further supported by the presence of post-tectonic Early Mesoproterozoic (c. 1600 Ma) felsic volcanics and granitoids in both areas (Peucat et al., 2002).

The rocks from the Wilkes Province also have direct correlatives in southern Australia (Duebendorfer, 2002; Bodorkos and Clarke, 2004). Rocks from the Wilkes Province and the Albany-Fraser region of Western Australia both record two periods of upper-amphibolite to granulite facies orogenesis. These events are dated between 1340–1290 Ma and 1210–1150 Ma (Black et al., 1992; Sheraton et al., 1992, 1993; Clark et al., 2000; Möller et al., 2002; Bodorkos and Clarke, 2004). In the eastern exposures of the Wilkes Province (Windmill Islands), the 1210–1150 Ma event overprints rocks that preserve evidence of the earlier 1340–1290 Ma event. The 1210–1150 Ma event is also recognised further to

the west in the Bunger Hills, although at this locality evidence for 1340–1290 Ma orogenesis is absent, and 1210–1150 Ma tectonism overprints both Archaean (c. 2640 Ma) and Palaeoproterozoic (c. 1700 Ma) felsic orthogneiss (Sheraton et al., 1992).

The Ruker Province (Kamenev et al., 1993; Mikhalsky et al., 2001) defines the final area of exposed basement belonging to the East Antarctic Craton (Fig. 1). It crops out in the upper catchment of the Lambert Glacier in the southern Prince Charles Mountains (Fig. 2a). The Ruker Province extends as far north as the central Mawson Escarpment on the eastern side of the Lambert Glacier where it is in sheared contact (Fig. 3) with the appreciably younger Lambert Province (Boger et al., 2001; Boger and Wilson, 2005). On the western side of the Lambert Glacier, the extent of both of the Ruker and Lambert Provinces are not clearly delineated. Previous studies, based primarily on the early Australian and Russian observational data, grouped all of the rocks south of Mt Johns and west

Fig. 2. (a) Generalised map of the Lambert basin illustrating the distribution of exposed rocks and the boundaries of the major tectonic Provinces. Ruker Province = Archaean granitoids (Mawson Suite) interleaved with Archaean (Menzies Series) and Proterozoic (Ruker and Sodrzhestvo Series) metasediments. Fisher Province = Mesoproterozoic (c. 1300 Ma) metavolcanics and intrusions. Rayner Province = high-grade Neoproterozoic (990–900 Ma) metamorphics. Lambert Province = Palaeo- to Neoproterozoic gneisses overprinted by middle to high-grade Early Cambrian (540–490 Ma) orogenesis. Davis and Mawson refer to permanent Australian stations. (b) Enlargement of boxed area in (a). Figure shows the main nunataks and the distribution of Archaean basement rocks (Mawson Suite and Menzies Series) and younger cover sequences in the southern Prince Charles Mountains (after Phillips unpublished data). Study area was within the Ruker Complex exposed along the southern Mawson Escarpment.

Fig. 3. Geological map (top) and cross-section (bottom) of the southern Mawson Escarpment. Numbers in legend are: (1) Archean granite gneiss; (2) early Cambrian granite gneiss (c. 550 Ma); (3) undifferentiated biotite and/or hornblende bearing gneiss; (4) marble and calc-silicate gneiss; (5) mafic gneiss and (6) quartzite and metapelitic schist. The trend and plunge of the major fold axes are shown on the map. Solid fold axes = folds of Archean age (F₃ in the text), dashed fold axes = folds of Early Cambrian age (Boger and Wilson, 2005). Areas A and B refer to the detailed cross-sections shown in Fig. 4.

of the Lambert Glacier (Fig. 2) as part of the Ruker Province (Kameniev et al., 1993; Mikhalsky et al., 2001). This broad region was subdivided into the Mawson granitoid suite, and three distinct sedimentary groups, the Menzies, Ruker and Sodruzhestvo Series. All were thought to have been deposited/intruded during the Late Archaean and the Mesoproterozoic, and subsequently deformed during the Neoproterozoic (Mikhalsky et al., 2001). As we will describe below, an entirely Archaean history is preserved in the rocks from the Mawson Suite and the Menzies Series, while recently acquired detrital zircon data (Phillips, unpublished data) from a number of nunataks west of the Lambert Glacier suggest that the remaining metasedimentary sequences post-date the Archaean intrusive and orogenic events recognised in these rocks. The Ruker and Sodruzhestvo Series therefore form cover sequences to what we consider the “Ruker Complex”—those rocks which were deformed and underwent middle- to high-grade metamorphism during the Archaean. On the basis of rock type and metamorphic grade, the rocks exposed at Mounts Rymill, Stinear, Menzies, Bayliss, Scherger and McCauley are also tentatively correlated with those

grouped here as the Ruker Complex (Fig. 2). The broad geographic distribution of the “Ruker Complex” and the undifferentiated cover sequences as defined by the new data of Phillips (unpublished) is illustrated in Fig. 2.

Within the Ruker Complex, the Mawson granitoid suite is generally interpreted to form an orthogneissic basement onto which the Menzies metasediments were deposited (Mikhalsky et al., 2001). Rocks of the Mawson Suite have been dated using a number of isotopic systems and have yielded ages from the Middle to Late Archaean. Summarising early Australian and Russian data, Mikhalsky et al. (2001) reported Rb–Sr isochron ages of 2762 ± 222 Ma and 2708 ± 90 Ma from orthogneissic rocks exposed along the southern Mawson Escarpment. Zircon thermal ionisation and Sm–Nd mineral-whole rock isochron ages from the same rocks yielded older ages. From two samples the Sm–Nd system yielded ages of 3124 ± 130 Ma and 3176 ± 140 Ma. U–Pb zircon ages between 3070 Ma and 2980 Ma were obtained from equivalent samples. These results are similar to those from Mt Bayliss where a granitic orthogneiss has given

an imprecise Rb–Sr isochron age of 2750 ± 402 Ma and from Mt Ruker where another granite gave a U–Pb zircon age of 3005 ± 57 Ma (Mikhalsky et al., 2001).

The oldest of the sedimentary cover sequences, the Menzies Series, is characterised by high-Cr, white to greenish, fuchsitic quartzite and the presence of kyanite in more aluminous rocks. Other common rock types include metaconglomerate, a variety of calc-silicates and marbles, and subordinate mafic to ultra-mafic schists

(Mikhalsky et al., 2001). The high-Cr content of the quartzites and an Rb–Sr muscovite age of c. 2580 Ma (Tingey, 1991) from a post-tectonic pegmatite (Mt Stinear) imply the Menzies Series was deposited during the Archaean. Coexisting kyanite, garnet and staurolite, suggest these rocks experienced mid- to high-pressure upper amphibolite facies metamorphic conditions. The eastern sectors of the Menzies Series (Mounts Rymill, Stinear and the southern Mawson Escarpment) are cross-cut by multiple generations of mostly undeformed mafic

Fig. 4. Structural cross-sections through the southern Mawson Escarpment. (a–e) Refer to area specific structural data. North is up for all stereonets. (a) Steep-dipping zone at the northern end of McCue Bluff. (b) McCue Bluff antiform (F_3). (c) Steep-dipping zone at the southern end of McCue Bluff including the exposures along the northern flanks of the Tingey Glacier. (d) Steep-dipping zone at the northern end of Rimmington Bluff including the exposures along the southern flanks of the Tingey Glacier. (e) Rimmington Bluff antiform (F_3). Best fit great circle in (a) defines the dominant $S_{0/1}$ trend in this region. Contoured plots in (c) and (d) are of poles to $S_{0/1}$. Best fit great circles define the dominant $S_{0/1}$ trend in each of these regions. Sample locations (diamonds) with ages are shown. Numbered stars refer to the locations of the photographs shown in Fig. 5.

dykes. The distribution of these dykes to the west remains unknown.

The overlying Ruker and Sodruzhestvo Series consist of greenschist to sub-greenschist facies sedimentary and metavolcanic rocks (Fig. 2). Although deformed, detrital zircons obtained from both of these sequences (Phillips unpublished data) suggest that this event (or events?) significantly post-date the intrusive and orogenic events recorded by the Mawson Suite and Menzies Series.

3. The southern Mawson Escarpment

The area of study is focused along the southern Mawson Escarpment (Fig. 2), a region that consists of interleaved granite orthogneiss of the Mawson Suite and strongly deformed layered and massive gneisses and metasediments of the Menzies Series. The gross structure of this region is defined by long wavelength (10–20 km), open, upright antiforms that are separated from each other by steep dipping high-strain zones (Fig. 3). Both the granitic orthogneiss and the metasediments contain concordant mafic to ultramafic bodies that show variable degrees stretching and boudinage. Characteristic of this region are large, cross cutting, and mostly undeformed amphibolised tholeiitic dykes.

3.1. Rock types

3.1.1. Mawson granitoid suite

Felsic orthogneisses of the Mawson granitoid suite constitute approximately 50% of the study area. These rocks crop out prominently at Rimmington and McCue Bluffs where they are exposed in the core of two regional antiforms and at the northern tip of Rimmington Bluff where a 1 km thick slab of granite orthogneiss is tectonically interleaved with massive hornblende gneiss and layered hornblende-biotite gneiss (Fig. 4). The Mawson granitoids exposed along the southern Mawson Escarpment are grey to brown in colour, strongly foliated, and homogeneous in appearance. Abundant foliation-parallel horizons of amphibolite produce a layered appearance at some localities (Fig. 5a). These rocks contain prominent, generally rounded, K-feldspar augen that are enclosed in a matrix of quartz, plagioclase, K-feldspar, hornblende and biotite. Garnet also occurs at some localities. The felsic orthogneisses are of granitic to trondhjemitic composition, metaluminous, and characterised by elevated concentrations of high field-strength (HFSE) and light rare-earth (LREE) elements (Sheraton et al., 1996; Mikhalsky et al., 2001). These characteris-

tics are reported from equivalent orthogneisses exposed sporadically throughout the southern Prince Charles Mountains and taken to suggest all of these granitoids are from a single suite (Sheraton et al., 1996; Mikhalsky et al., 2001). Leucosomes of predominantly quartz and K-feldspar are developed within the granitic gneisses particularly in the north of the study area (see sample 9828-189 below). These provide some evidence that these granites were re-melted subsequent to their emplacement.

3.1.2. Massive hornblende gneiss

Also of probable orthogneissic origin, massive hornblende gneiss crops out as thick (typically a kilometre or less) layers tectonically interleaved with both felsic orthogneiss and rocks of metasedimentary origin (Fig. 4). These rocks are massive, mostly homogeneous and finer-grained than the rocks of the Mawson Suite. They are grey in colour and darker in appearance than the Mawson orthogneisses owing to their relatively high proportion of hornblende (30–50%). Other rock forming minerals include quartz, plagioclase and K-feldspar. Sphene occurs as a common accessory. Coarse-grained K-feldspar is found in the matrix in some samples and is common as a major constituent in leucosomes. These are occasionally present and, in addition to K-feldspar and quartz, commonly contain clinopyroxene and coarse-grained hornblende that can both be enclosed by large garnet porphyroblasts. Garnet is also locally present in the matrix.

3.1.3. Pelitic to psammopelitic schist

Pelitic and psammopelitic schists crop out as relatively narrow layers interleaved with quartzite and layered hornblende-biotite gneiss (Fig. 4). The more pelitic of these rocks contain the mid- to high-pressure Barrovian index minerals garnet, staurolite and kyanite. These minerals normally occur as porphyroblasts in a finer-grained muscovite and quartz matrix. The matrix may or may not also include biotite and, in some rocks, chlorite. Sillimanite and cordierite can also be found in some rocks. A simpler assemblage of garnet, biotite, muscovite and quartz characterises rocks of more psammopelitic composition. All of the pelitic to psammopelitic rocks contain a well-developed schistose foliation (S_1) defined by a preferred alignment of mica. In the more pelitic layers, kyanite and staurolite are aligned parallel with this foliation. In contrast, garnet is wrapped by the foliation and would appear to pre-date staurolite and kyanite. These rocks vary from a yellowish brown to grey colour depending on mineral assemblage.

Fig. 5. Structural features of the Ruker Province. (a) D₁ high-strain zone—Rimmington Bluff (geologists circled lower left for scale). (b) Isoclinal F₂ fold—north end of Rimmington Bluff. (c) Steep-dipping D₃ high-strain zone—north end of McCue Bluff. High-strain zone is the well-foliated region to the left of the dashed line (height of cliff is approximately 1000 m). (d) Steeply-dipping D₃ high-strain zone—south end of McCue Bluff. Dashed lines highlight the dragging of the S₁ foliation into the high-strain zone. Dragging shows north-up (left-up) sense of transport. (e) Outcrop scale retrograde mylonites with minor offset—Tingey Glacier. (f) Regional scale retrograde mylonite—Rimmington Bluff. Mylonite separates massive hb + bi gneiss (left) from interlayered quartzite and metapelite (right). Dashed line marks the mylonite plane (geologists circled lower left for scale).

3.1.4. Layered felsic to intermediate gneiss

Layered felsic to intermediate gneisses crop-out extensively in the valley of the Tingey Glacier and at both the northern and southern-central sections of McCue Bluff (Fig. 4). This rock type is also known from the southern end of Rimmington Bluff. These rocks are layered on the centimetre to metre scale and consist of quartz and K-feldspar rich layers that separate more mafic layers dominated by green-brown hornblende and/or biotite. Biotite and, less commonly, garnet are also present in the hornblende dominated layers. Compositionally intermediate layers are also present. These consist of more disseminated hornblende ($\pm$ biotite, $\pm$ garnet) and in some instances clinopyroxene distributed in a quartz, K-feldspar and/or plagioclase matrix.

3.1.5. Mafic gneiss

Mafic gneisses appear black and either coarsely layered or massive in outcrop. They form relatively limited continuous exposures through McCue Bluff (Fig. 4) and appear elsewhere as large boudins or discontinuous layers within other rock types. These rocks commonly contain relic clinopyroxene, surrounded by green-brown hornblende and less commonly biotite. Plagioclase and some quartz are found in the matrix. Other layers lack clinopyroxene and are compositionally similar to the ferro-magnesium layers observed within the layered felsic to intermediate gneiss described above. These may also contain garnet. The larger packages of mafic gneiss are locally interleaved with intermediate clinopyroxene, plagioclase and quartz gneiss as well as pyroxenite.

3.1.6. Post-tectonic dolerite dykes

Post-tectonic dolerite dykes are prominent cross cutting intrusions in the study area. They are mostly of typical olivine or quartz tholeiite composition (Mikhalsky et al., 2001, and references therein) and are variably amphibolised. The most pristine dykes crop out toward the northern end of McCue Bluff. These contain igneous clinopyroxene and plagioclase and an opaque implied to be ilmenite. More commonly, clinopyroxene is partly replaced by green-brown hornblende and to a lesser degree by biotite, while the opaque is rimmed by sphene. In the more southerly exposed dykes, this process of amphibolization has gone to completion. These rocks contain equirranular hornblende, biotite, sphene, quartz and plagioclase.

3.2. Structural geology

The earliest fabric element recognised along the southern Mawson Escarpment is an intense layer-parallel foliation (S_1). This foliation defines the dominant fabric preserved in both the orthogneissic and in the intercalated metasedimentary rocks. In the latter rocks S_1 is parallel with bedding and defines a composite structural/lithological layering ($S_{0/1}$). S_1 is defined by aligned biotite, hornblende, quartz and deformed feldspar augen in the intrusive rocks, and by a strong preferred orientation of muscovite, biotite and kyanite in the metasedimentary rocks. The intensity of S_1 is variable. Distinct D_1 high-strain zones, commonly up to hundreds of metres in width, are observed throughout the studied section (Fig. 5a). Within these high-strain zones deformation appears to have been constrictional and produced an intense lineation (L_1) on more diffuse S_1 surfaces (L–S tectonite). Outside the high-strain zones L_1 is also well-developed and is everywhere parallel to the axes of F_2 folds. Kinematic indicators were observed locally, although these mostly gave vergence to the next F_2 closure. Consequently we were not able to define a consistent direction of D_1 tectonic transport. F_1 folds were not recognised.

F_2 folds form isoclinal layer-parallel closures that are recognisable on both the outcrop (Fig. 5b) and map scale (Fig. 4). They define the dominant mesoscale structures observed along the southern Mawson Escarpment. The orientation of F_2 fold axes is variable. On the southern side of the Tingey Glacier, F_2 folds plunge moderately to the northwest (Fig. 4). To the north of the Tingey Glacier, F_2 fold axes distribute within the S_1 plane. This difference in F_2 morphology is interpreted to reflect the development of both cylindrical and non-cylindrical folds, which in turn is interpreted to be a function of

localised variations in D_2 strain. Unfolding the effects of D_3 south of the Tingey Glacier returns a mean F_2 fold axis to a northwest trend and sub-horizontal plunge. Although L_1 is everywhere well developed, we interpret the parallelism of L_1 and F_2 to suggest L_1 is aligned parallel with the vorticity vector (i.e. parallel with the axis of rotation) rather than the direction of tectonic transport. This is not an uncommon phenomenon when deformation deviates from ideal simple shear (Passchier, 1997). With this in mind we assume tectonic transport was orthogonal to the trend and plunge of F_2 (and thus also L_1) and consequently imply a NE–SW direction of transport for D_2 . The non-cylindrical nature of the F_2 folds north of the Tingey Glacier prohibits the estimation of a meaningful direction of D_2 transport from this region. The extreme attenuation of F_2 folds (Fig. 5b), their localised non-cylindrical nature and the kilometre scale intercalation of granitic orthogneiss and metasediments is suggestive of very high D_2 strain rates. We interpret this to reflect the likely development of these structures via thrust tectonics. A similar conclusion was drawn by the previous Russian studies from this area (Kamenev, 1991).

Subsequent D_3 deformation buckled the terrane into a series of macroscale upright antiforms with no evidence for the development of equivalent mesoscale structures. In the study area two major F_3 antiforms with wavelengths of approximately 10 km were observed (Fig. 4). These folds plunge shallowly to the east (McCue Bluff) and west (Rimmington Bluff). An axial planar fabric is discernible locally, but is generally poorly developed. Steep north-dipping and east-west trending high-strain zones (Fig. 5c) truncate both limbs of the F_3 antiform exposed at McCue Bluff. These high-strain zones are orientated parallel with the F_3 fold axial plane (Fig. 4) and are taken to represent localised zones of more intense D_3 strain. The D_3 high-strain zones are characteristically poorly lineated and are interpreted to represent S–L tectonites. They rotate the adjacent foliation and offset mafic units to show reverse, north-up movement (Fig. 5d). Given the orientation of both F_3 folds and D_3 high-strain zones, D_3 is considered to have formed in response to north-south directed compression. Kinematic asymmetry (Fig. 5d) points to north-up or top to the south directed transport.

Ductile features attributed to D_1 – D_3 are cross cut by a series of largely undeformed mafic dykes (Fig. 5c) and by discrete low-grade mylonites. The mafic dykes pre-date the mylonites and intrude in four dominant orientations striking approximately north, east-north-east, east and southeast. The east-west trending dyke set appears to be the oldest and is chemically distinct from the remaining

Table 1
U–Pb–Th isotopic compositions of zircons from the Mawson granitoid suite

Grain spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ (%)	Radiogenic ratios						Age (Ma)						
							²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²³⁵ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²³⁵ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	Disc (%)
Sample 9828-101 (Rimmington Bluff granite)																			
1.1	370	231	0.62	165	0.00002	0.03	0.6219	0.010	21.31	0.35	0.2485	0.001	3117	38	3153	16	3175	7	98
1.2	317	168	0.53	135	0.00009	0.11	0.6047	0.011	20.76	0.41	0.2490	0.001	3049	44	3128	19	3179	9	96
2.1	195	88	0.45	82	0.00011	0.14	0.6027	0.016	20.58	0.58	0.2476	0.002	3041	65	3119	28	3170	12	96
3.1	404	164	0.41	169		<0.01	0.6081	0.009	21.09	0.34	0.2515	0.001	3062	35	3143	16	3194	9	96
4.1	202	116	0.57	87		<0.01	0.6068	0.012	20.83	0.48	0.2490	0.002	3057	50	3131	22	3178	12	96
4.2	309	175	0.57	131	0.00008	0.10	0.5994	0.009	20.28	0.37	0.2454	0.002	3028	38	3105	18	3155	12	96
5.1	827	83	0.10	233	0.00001	0.01	0.4499	0.006	12.63	0.20	0.2037	0.001	2395	26	2653	15	2856	11	84
5.2	234	133	0.57	101	0.00015	0.19	0.6133	0.012	20.40	0.43	0.2412	0.002	3083	47	3110	21	3128	12	99
6.1	979	540	0.55	413	0.00002	0.03	0.5970	0.008	20.48	0.28	0.2488	0.001	3018	31	3115	13	3177	5	95
7.1	169	82	0.48	75	0.00012	0.16	0.6375	0.013	21.82	0.48	0.2482	0.002	3179	50	3176	22	3173	12	100
8.1	1957	43	0.02	388	0.00002	0.02	0.3330	0.004	7.84	0.10	0.1709	0.001	1853	18	2213	11	2566	6	72
8.2	212	135	0.64	96	0.00009	0.11	0.6313	0.016	21.59	0.60	0.2480	0.002	3155	64	3165	27	3172	13	99
Sample 9828-039 (Rimmington Bluff)																			
1.1	163	97	0.60	106	0.00003	0.04	0.6512	0.017	22.59	0.60	0.2516	0.001	3233	66	3209	26	3195	8	101
2.1	188	95	0.51	121	0.00002	0.03	0.6529	0.014	22.52	0.52	0.2502	0.001	3239	56	3207	23	3186	9	102
3.1	237	125	0.52	151	0.00005	0.06	0.6457	0.010	22.01	0.38	0.2472	0.001	3212	40	3184	17	3167	9	101
4.1	283	150	0.53	179	0.00007	0.09	0.6509	0.010	20.84	0.33	0.2322	0.001	3232	38	3131	15	3067	6	105
4.2	595	287	0.48	375	0.00000	0.00	0.6422	0.008	21.83	0.31	0.2465	0.001	3198	33	3176	14	3162	6	101
5.1	199	89	0.45	125	0.00005	0.06	0.6510	0.010	21.69	0.36	0.2416	0.001	3232	39	3170	16	3131	8	103
5.2	581	12	0.02	236	0.00001	0.01	0.4845	0.007	11.47	0.20	0.1717	0.001	2547	32	2562	16	2575	11	99
6.1	62	20	0.32	41	0.00013	0.16	0.6962	0.020	24.33	0.73	0.2535	0.002	3406	75	3282	30	3207	12	106
6.2	439	171	0.39	275	0.00001	0.01	0.6521	0.009	22.14	0.33	0.2463	0.001	3236	35	3190	14	3161	5	102
7.1	148	67	0.46	95	0.00005	0.06	0.6627	0.015	22.71	0.54	0.2485	0.001	3278	58	3215	23	3175	9	103
7.2	527	9	0.02	212	0.00002	0.03	0.4796	0.007	11.48	0.18	0.1736	0.001	2525	29	2563	15	2593	9	97
8.1	185	91	0.49	121	0.00003	0.04	0.6635	0.011	22.63	0.39	0.2473	0.001	3281	41	3211	17	3168	8	104
8.2	656	100	0.15	332	0.00003	0.04	0.5620	0.007	16.90	0.24	0.2181	0.001	2875	30	2929	14	2967	7	97
8.3	544	6	0.01	204	0.00004	0.06	0.4471	0.007	10.84	0.17	0.1759	0.001	2382	29	2510	15	2615	8	91
9.1	245	119	0.49	156	0.00003	0.04	0.6505	0.011	22.23	0.40	0.2479	0.001	3230	43	3194	17	3171	7	102
10	213	107	0.51	133	0.00002	0.02	0.6353	0.014	21.89	0.53	0.2499	0.002	3171	56	3179	24	3184	11	100
10	735	14	0.02	268	0.00001	0.02	0.4352	0.006	10.46	0.16	0.1743	0.001	2329	29	2476	14	2599	6	90
11	167	73	0.44	106	0.00003	0.04	0.6531	0.012	22.64	0.47	0.2514	0.002	3240	47	3211	20	3194	12	102
11	403	50	0.12	162		<0.01	0.4661	0.008	11.40	0.23	0.1774	0.001	2466	37	2556	19	2628	12	94
Sample 9828-190 (McCue Bluff granite-south)																			
1.1	715	90	0.13	277	0.00002	0.02	0.6014	0.008	19.78	0.29	0.2386	0.001	3036	33	3081	14	3110	7	98
1.2	267	159	0.60	135	0.00010	0.13	0.6923	0.013	27.13	0.54	0.2842	0.001	3391	49	3388	20	3386	7	100
2.1	817	36	0.04	260	0.00001	0.01	0.5192	0.008	14.37	0.23	0.2007	0.001	2696	34	2774	15	2832	6	95
2.2	592	217	0.37	288	0.00005	0.06	0.6958	0.010	27.06	0.42	0.2821	0.001	3405	39	3386	15	3374	6	101

3.1	411	71	0.17	170	0.00001	0.01	0.6329	0.009	21.62	0.34	0.2478	0.001	3161	35	3167	15	3170	9	100
3.2	651	894	1.37	357	0.00004	0.05	0.6559	0.009	25.38	0.37	0.2807	0.001	3251	36	3323	14	3367	5	97
4.1	360	61	0.17	135	0.00004	0.06	0.5782	0.011	18.73	0.39	0.2349	0.002	2941	44	3028	20	3086	11	95
4.2	527	434	0.82	270	0.00001	0.01	0.6740	0.011	25.85	0.45	0.2782	0.001	3321	42	3341	17	3353	8	99
5.1	526	65	0.12	205	0.00006	0.07	0.6010	0.009	20.30	0.39	0.2450	0.003	3034	35	3106	19	3153	17	96
5.2	423	250	0.59	191	0.00010	0.12	0.6298	0.011	22.45	0.41	0.2585	0.001	3149	42	3203	18	3238	9	97
6.1	1251	193	0.15	674	0.00007	0.09	0.6016	0.007	20.40	0.26	0.2459	0.001	3036	30	3110	13	3159	4	96
6.2	317	146	0.46	220	0.00003	0.04	0.7105	0.010	27.65	0.44	0.2823	0.001	3460	39	3407	16	3376	8	103
7.1	174	42	0.24	96	0.00002	0.02	0.6115	0.013	19.67	0.53	0.2333	0.003	3076	50	3075	26	3075	24	100
7.2	323	86	0.27	209	0.00055	0.67	0.6829	0.013	26.24	0.79	0.2786	0.006	3356	52	3355	30	3355	33	100
8.1	352	77	0.22	200	0.00004	0.05	0.6287	0.009	21.25	0.33	0.2452	0.001	3144	36	3150	15	3154	7	100
9.1	432	44	0.10	226	0.00001	0.01	0.5993	0.009	19.23	0.32	0.2327	0.001	3027	37	3053	16	3071	8	99
9.2	314	150	0.48	204	0.00003	0.04	0.6686	0.011	24.98	0.46	0.2710	0.002	3300	43	3307	18	3312	9	100
10	301	54	0.18	170	0.00064	0.81	0.6252	0.010	21.48	0.40	0.2492	0.002	3130	39	3161	18	3180	13	98
10	215	102	0.47	143	0.00006	0.07	0.6822	0.015	25.90	0.63	0.2754	0.002	3353	59	3343	24	3337	10	101
Sample 9828-190 (McCue Bluff granite-north)																			
1.1	1405	101	0.07	732	0.00054	0.67	0.5928	0.007	19.95	0.27	0.2441	0.001	3001	30	3089	13	3147	6	95
2.1	2483	479	0.19	1397	0.00643	8.08	0.6309	0.008	20.97	0.34	0.2411	0.002	3153	31	3137	16	3127	14	101
3.1	917	86	0.09	536	0.00002	0.02	0.6601	0.009	22.81	0.33	0.2507	0.001	3268	35	3219	14	3189	6	103
4.1	1007	91	0.09	412	0.00003	0.04	0.4670	0.111	15.53	3.72	0.2412	0.002	2470	508	2848	259	3128	16	79
4.2	1161	136	0.12	587	0.00008	0.10	0.5713	0.008	19.14	0.28	0.2430	0.001	2913	32	3049	14	3140	5	93
5.1	1604	167	0.10	896	0.00013	0.16	0.6308	0.008	21.88	0.32	0.2516	0.001	3153	34	3179	14	3195	6	99
5.2	3184	508	0.16	1790	0.00011	0.14	0.6283	0.007	21.38	0.26	0.2467	0.000	3143	29	3156	12	3164	3	99
6.1	1135	112	0.10	587	0.00006	0.08	0.5851	0.008	19.86	0.28	0.2462	0.001	2969	31	3085	14	3160	6	94
7.1	920	91	0.10	493	0.00010	0.12	0.6066	0.009	20.47	0.32	0.2448	0.001	3056	35	3114	15	3151	7	97

Uncertainties are given at the 1σ level; $f_{206\%}$ denotes the percentage of ^{206}Pb that is common Pb; for % Disc, 100 denotes a concordant analysis.

three dyke sets (Mikhalsky et al., 2001). Following emplacement, the mafic dykes were dissected, or where appropriately orientated, reworked and retrogressed to greenschist facies by a subsequent phase of deformation localised along discrete mylonites. These mylonites vary from discrete millimetre scale zones showing minor offset (Fig. 5e) to large retrogressed metre scale shear zones with kilometre scale offsets (Fig. 5f). Although variable, a dominant east trend and moderate northerly dip was observed for these structures. Both reverse and normal offsets were observed (Fig. 4).

3.3. Geochronology

Zircons for SHRIMP analysis were separated by standard heavy liquid and magnetic procedures, then by hand picking. They were then mounted in epoxy resin discs along with fragments of zircon standard SL13. The

discs were polished and Au-coated before being analysed on the SHRIMP I or SHRIMP II ion-microprobe housed at the Australian National University. Cathodoluminescent (CL) imaging was conducted to assess the internal structure of the unknown zircons. A primary beam of O^- ions was used to sputter positive secondary ions from areas $\sim 25 \mu\text{m}$ in diameter from individual sectioned zircons. Zr_2O^+ , $^{206}Pb^+$, $^{207}Pb^+$, $^{208}Pb^+$, $^{238}U^+$, $^{232}ThO^+$, and $^{238}UO^+$, were measured in cycles by magnetic field switching, seven cycles per data set. Analysis of unknowns was interspersed with analyses of the standard SL13 (Compston et al., 1984, 1992) in order to monitor the differential fractionation between U and Pb. Data reduction and analysis was undertaken using the SQUID and ISOPLOT software of Ludwig (2001). Reported ages are based on radiogenic $^{207}Pb/^{206}Pb$ with errors given with 2σ confidence limits.

Fig. 6. Sample 9828-101 (Mawson Suite; Rimmington Bluff). (a) Cathodoluminescent response of representative zircon grains with location and age of individual analyses shown. (b) Wetherall concordia plot of SHRIMP data. Confidence limits for ages and error ellipses are 2σ and 1σ , respectively. White error ellipses are discordant or suspect data that are not used to define weighted mean age.

3.3.1. Pre-tectonic granite gneiss

Isotopic data was collected from four samples from the Mawson Suite. Samples were collected from the northern end of the Rimmington Bluff (sample 9828-039), at the northern tip of the main Rimmington Bluff granite (sample 9828-101), and at both the northern (sample 9828-210) and southern (sample 9828-190) ends of the McCue Bluff granite. The location of these samples is illustrated in Fig. 4.

3.3.1.1. Rimmington Bluff granite (south). Sample 9828-101 was collected from the northern margin of the main Rimmington Bluff granite. This rock is a biotite and hornblende bearing granitoid that crops out in the core of the F₃ antiform that runs through the central section of Rimmington Bluff (Fig. 4). Zircons from this sample are pale brown, translucent, and display mostly euhedral grain shapes. Most grains are approximately 100–200 μm in length and have average elongation ratios

of 2–3:1. Zircons contain between 200 and 1000 ppm U and have relatively constant Th/U ratios of approximately 0.5 (Table 1). Cathodoluminescent (CL) images show that most zircons have distinct oscillatory zoning (Fig. 6a). Dark rims are found on some grains, although these tend to parallel the earlier more luminescent bands and are interpreted to represent the last phase of zircon growth rather than a younger phase of new zircon.

Twelve analyses of eight zircon grains were obtained from both core and rim positions within the analysed zircons. Ten of the analyses produced a single age population from which a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3177 ± 6 Ma (MSWD=0.92) was obtained. The remaining two analyses were discordant and, together with the other mostly concordant analyses, define upper and lower intercept ages of 3197 ± 25 Ma and 1343 ± 160 Ma (MSWD=4.4). Given the characteristically igneous zoning observed and the straightforward age population obtained, the weighted mean age of

Fig. 7. Sample 9828-039 (Mawson Suite; Rimmington Bluff). (a) Cathodoluminescent response of representative zircon grains with location and age of individual analyses shown. (b) Wetherall concordia plot of SHRIMP data. Confidence limits for ages and error ellipses are 2σ and 1σ , respectively. White error ellipses are discordant or suspect data that are not used to define weighted mean ages.

3177 ± 6 Ma is interpreted to date the emplacement of the Rimmington Bluff granite.

3.3.1.2. Rimmington Bluff granite (north). Sample 9828-039 was obtained from a 1 km thick layer of granite gneiss that crops out at the northern end of Rimmington Bluff (Fig. 4). The sample was collected from within the 50 m wide D₂ high-strain zone observed at this locality (Fig. 5a). This high-strain zone trends east and dips moderately to the north. Compared to the previous sample, this granite was much more intensely deformed and contained garnet as opposed to hornblende as the primary mafic mineral.

Zircons from sample 9628-039 are translucent and colourless to pale yellow in colour. They form a population of mostly euhedral zircons that vary in length from 100 to 400 μm . They have length to width ratios of approximately 2–3:1 and, similar to sample 9828-101, show well-developed oscillatory zoning (Fig. 7). However, some grains show a narrow unzoned low-CL rim that truncates the oscillatory zoning observed in the core (Fig. 7a). U concentrations of between 100 and 200 ppm are observed for the zircon cores and contrast with the low-CL rims where U concentration was in the order of 500 ppm (Table 1). Higher Th/U ratios are also observed for the zircon cores (0.30–0.60) when compared to the zircon rims (0.02–0.10).

Nineteen zircon analyses were obtained from this sample. Thirteen analyses were of zircon cores and six analyses were of zircon rims. Eleven of the data from the zircon cores formed a concordant to slightly reversely discordant population yielding a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3174 ± 9 Ma (Fig. 7b). Given the oscillatory zoning observed in the cores, this age is interpreted to date the emplacement of the granite. Five of the six rim analyses formed a slightly discordant population with ages near c. 2600 Ma. When a line is forced through the same lower intercept (c. 1350 Ma) as that obtained from sample 9828-101, an upper intercept age of 2691 ± 100 Ma is obtained. Ignoring two of the data, an identical but statistically better age of 2752 ± 30 Ma (MSWD = 1.04) is obtained. Both ages are within error of those obtained from samples 9828-189 and 9828-218. These two samples, described below, specifically date orogenesis and regional metamorphism. With this in mind, the upper intercept age (2752 ± 30 Ma) is taken as the best estimate of the age of the low-CL zircon rims. The relatively high concentrations of U, low Th/U, and the structureless nature of these rims is consistent with a metamorphic origin for this phase of zircon. The remaining rim analysis gave an age of 2967 ± 7 Ma a result approximately half way between the age obtained from

the zircon cores and that obtained from the majority of the rims. This point is interpreted to represent a mixed analysis and is not considered significant.

3.3.1.3. McCue Bluff granite (south end). Sample 9828-190 was collected from the southern end of the McCue Bluff granite (Fig. 4). This granite is similar in appearance to the Rimmington Bluff granite, although the mineral assemblage lacked hornblende. Zircons from this gneiss are translucent and pale brown. They form a subhedral population that have elongation ratios are approximately 3 or 4 to 1. Grain lengths are typically between 200 and 400 μm . The majority of zircons contain 200–500 ppm U, although grains with higher U concentrations (c. 1000 ppm) were also present (Table 1). The observed Th/U ratios were between 0.1 and 0.8. In contrast to the previous samples, two stages of oscillatory zoned zircon was observed. The earliest generation was observed in the cores of some, but not all, zircon grains (Fig. 8a). Where observed, this oldest stage of zircon consisted of both complete cores and broken fragments. For the fragmented cores, the boundaries of this and the subsequent phase of zircon growth is marked by sharply truncated edges overgrown by an obvious phase of concentrically zoned new zircon of similar appearance (Fig. 8a). For the unbroken cores, the boundaries between the different generations of zircon are commonly concordant and more difficult to identify. However, a distinct band of low-CL zircon commonly marks the onset of the second phase of zircon growth (Fig. 8a), a feature that allowed the two generations of zircon to be distinguished. In some instances, the core/rim boundaries are also visible in transmitted light. Both generations of zircon are considered to be of magmatic origin. The older zircon is interpreted to be of xenocrystic origin and to be inherited from the source of these granites. The younger rims are inferred to record the emplacement of the McCue Bluff granite. The low-CL rims observed in the previous sample were not present.

Nineteen analyses of 10 zircon grains were obtained from sample 9828-190 (Table 1). Nine analyses were of the xenocrystic zircon cores while the remaining analyses were of the zoned overgrowths. Six of the nine core analyses yielded a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3370 ± 11 Ma (MSWD = 1.7). The remaining three yielded slightly younger ages of 3238 ± 9 Ma, 3312 ± 9 Ma and 3337 ± 10 Ma. Of the nine rim analyses, five yielded a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3160 ± 6 Ma (MSWD = 1.15). A further three can be inferred to produce a slightly younger age of 3076 ± 12 Ma (MSWD = 0.6). This latter age is anomalous when the emplacement ages of the other sample

Fig. 8. Mawson Suite samples from McCue Bluff. (a and c) Cathodoluminescent response of representative zircon grains from sample 9828-190 (a) and sample 9828-210 (c). Location and age of individual analyses shown. (b and d) Wetherall concordia plots of SHRIMP data for samples 9828-190 (b) and 9828-210 (d). Confidence limits for ages and error ellipses are 2σ and 1σ , respectively. White error ellipses are discordant or suspect data that are not used to define weighted mean ages.

from this granite (9828-210: 3162 ± 12 Ma), and the two ages obtained for the Rimmington Bluff granite (3177 ± 6 Ma and 3174 ± 9 Ma), are taken into consideration. Thus, we suggest these younger zircons may record subtle lead loss after an event at c. 3160 Ma rather than a stage of new zircon growth at c. 3080 Ma. When combined with a discordant analysis with an age of 2832 ± 6 Ma, all of the rim analyses define upper and lower intercept ages of 3194 ± 80 Ma and 2240 ± 420 Ma. The upper intercept age is within error of the age obtained from the older zircon rims. From these results it is concluded that the McCue Bluff granite melted from a Middle Archaean source (3370 ± 11 Ma) and crystallised at 3160 ± 6 Ma.

3.3.1.4. McCue Bluff granite (north end). Sample 9828-210 was collected at the northern end of McCue Bluff (Fig. 4). Similar to sample 9828-101 this rock was a biotite and hornblende bearing granitoid. Zircons from this sample are dark brown in colour and gen-

erally opaque. They have euhedral to subhedral grain shapes and form a population of 200–400 μm long zircons with average elongation ratios 2–3:1 (Fig. 8c). Relative to zircons from the other basement gneisses, zircons from this rock have high U concentrations (1000–3000 ppm) and low Th/U ratios (≤ 0.10). Zircons from this rock have a notably lower CL response and a more subdued internal structure when compared to those described above. Nevertheless oscillatory zoning was present (Fig. 8c) and, similar to the other samples, was taken to suggest an igneous origin of these zircons. These zircons also showed no obvious overgrowths and did not contain evidence of an inherited component.

Isotopic data was obtained from both core and rim localities from seven grains (Table 1). These data form a single but a somewhat scattered population that yielded a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3162 ± 13 Ma (MSWD=10.8; Fig. 8d). This age is interpreted to date the emplacement of the McCue Bluff granite and

confirms the inferred emplacement age of 3160 ± 6 Ma obtained from the previous sample.

3.3.2. Syn-tectonic leucosomes

Sample 9828-189 is obtained from a coarse-grained quartz, K-feldspar, plagioclase and biotite bearing leucosome sampled from within the McCue Bluff granite

near its southern margin (Fig. 4). The leucosome forms a mostly foliation parallel body that is folded by an outcrop scale F_2 fold (Fig. 9a). The leucosome contains no internal foliation and locally cross cuts S_1 observed within its granitoid host (Fig. 9a). Based on these leucosome/foliation relationships, partial melting and leucosome formation is interpreted to have post-dated the

Fig. 9. Syn-tectonic leucosomes from McCue Bluff. (a and d) Outcrop photographs of samples 9828-189 (a) and 9828-218 (d). (b and e) Cathodoluminescent response of representative zircon grains from samples 9828-189 (b) and 9828-218 (e). Location and age of individual analyses shown. (c and f) Wetherall concordia plots of SHRIMP data for 9828-189 (c) and 9828-218 (f). Confidence limits for ages and error ellipses are 2σ and 1σ , respectively. White error ellipses are discordant or suspect data that are not used to define weighted mean ages.

Table 2

U–Pb–Th isotopic compositions of zircons from the syn- and post-tectonic leucosomes and intrusions

Grain spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ (%)	Radiogenic ratios						Age (Ma)					
							²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁷ Pb/ ²³⁵ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	²⁰⁶ Pb/ ²³⁰ U	±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	Disc (%)	
Sample 9828-189 (syn-tectonic leucosome-McCue Bluff)																		
1.1	456	248	0.54	242	0.00004	0.05	0.5530	0.0074	14.93	0.22	0.1958	0.0009	2837	31	2792	7	102	
2.1	358	108	0.30	177	0.00007	0.10	0.5462	0.0077	14.64	0.22	0.1944	0.0007	2810	32	2780	6	101	
3.1	252	81	0.32	120	0.00017	0.23	0.5216	0.0079	14.17	0.23	0.1970	0.0010	2706	34	2802	8	97	
4.1	1556	60	0.04	601	0.00005	0.07	0.4575	0.0054	11.10	0.14	0.1760	0.0004	2428	24	2616	4	93	
5.1	377	146	0.39	185	0.00017	0.23	0.5299	0.0079	14.28	0.24	0.1955	0.0012	2741	33	2789	10	98	
6.1	338	125	0.37	168	0.00065	0.87	0.5411	0.0075	14.53	0.23	0.1948	0.0013	2788	31	2783	11	100	
7.1	274	101	0.37	129	0.00043	0.58	0.5125	0.0097	13.77	0.30	0.1949	0.0018	2667	42	2784	15	96	
8.1	1547	446	0.29	714	0.00007	0.10	0.5088	0.0067	13.72	0.19	0.1955	0.0005	2651	28	2789	4	95	
9.1	1877	222	0.12	968	0.00006	0.08	0.5698	0.0071	18.83	0.25	0.2397	0.0007	2907	29	3118	5	93	
10.1	368	155	0.42	179	0.00008	0.11	0.5218	0.0101	14.07	0.30	0.1956	0.0014	2707	43	2790	12	97	
11.1	907	338	0.37	478	0.00096	1.22	0.5560	0.0081	18.28	0.28	0.2385	0.0010	2850	33	3110	7	92	
Sample 9828-051 (syn-tectonic leucosome-McCue Bluff)																		
1.1	4117	126	0.03	2533	0.000003	0.00	0.5951	0.0063	15.93	0.17	0.1941	0.0004	3010	25	2777	3	108	
2.1	5439	86	0.02	3421	0.000002	0.00	0.6115	0.0065	16.20	0.18	0.1921	0.0003	3076	26	2761	3	111	
3.1	3853	121	0.03	2391	0.000000	0.00	0.6000	0.0062	16.03	0.17	0.1938	0.0002	3030	25	2775	1	109	
4.1	2367	102	0.04	1442	0.000003	0.00	0.5866	0.0066	15.85	0.18	0.1960	0.0002	2976	27	2793	2	107	
5.1	4934	148	0.03	3083	0.000002	0.00	0.6047	0.0064	16.13	0.18	0.1935	0.0003	3049	26	2772	2	110	
6.1	3174	115	0.04	1973	0.000002	0.00	0.6002	0.0062	16.08	0.17	0.1943	0.0002	3030	25	2779	2	109	
7.1	3715	21	0.01	2273	0.000003	0.00	0.5953	0.0066	15.89	0.19	0.1936	0.0005	3011	27	2773	4	109	
8.1	3575	251	0.07	2177	0.000002	0.00	0.5860	0.0062	15.70	0.17	0.1943	0.0003	2973	25	2779	2	107	
9.1	5580	102	0.02	3540	0.000002	0.00	0.6161	0.0064	16.35	0.17	0.1925	0.0002	3094	25	2763	2	112	
10.1	3835	119	0.03	2320	0.000001	0.00	0.5849	0.0095	15.67	0.27	0.1943	0.0006	2969	39	2779	5	107	
1.2	4132	88	0.02	2524	0.000004	0.01	0.5925	0.0061	15.80	0.17	0.1935	0.0002	2999	25	2772	2	108	
2.2	5803	38	0.01	3573	0.000003	0.00	0.5997	0.0063	15.88	0.17	0.1920	0.0004	3029	25	2760	4	110	
11.1	3362	89	0.03	2011	0.000083	0.11	0.5790	0.0060	15.45	0.17	0.1935	0.0003	2945	25	2772	3	106	
12.1	2851	61	0.02	1689	0.000003	0.00	0.5746	0.0060	15.33	0.17	0.1935	0.0004	2927	25	2772	3	106	
13.1	5428	42	0.01	3304	0.000002	0.00	0.5928	0.0062	15.70	0.17	0.1921	0.0004	3001	25	2761	4	109	
Sample 9828-051 (post-tectonic pegmatite-Rimington Bluff)																		
1.1	70	29	0.42	55	0.00015	0.19	0.6667	0.0227	22.78	0.83	0.2478	0.0024	3293	88	3171	16	104	
1.2	1314	47	0.04	629	0.00002	0.03	0.4694	0.0059	11.33	0.15	0.1750	0.0006	2481	26	2606	5	95	
2.1	54	25	0.47	42	0.00013	0.16	0.6623	0.0196	22.48	0.73	0.2462	0.0025	3276	76	3161	16	104	
2.2	670	47	0.07	330	0.00002	0.03	0.4771	0.0064	11.76	0.16	0.1788	0.0006	2515	28	2642	5	95	
3.1	57	23	0.40	42	0.00026	0.33	0.6288	0.0160	21.34	0.61	0.2461	0.0025	3145	64	3160	16	100	
3.2	912	41	0.05	471	0.00013	0.18	0.5058	0.0087	12.30	0.22	0.1764	0.0005	2639	37	2619	5	101	
4.1	497	32	0.06	260	0.00113	1.55	0.5041	0.0065	12.47	0.21	0.1794	0.0016	2631	28	2647	15	99	
5.1	535	38	0.07	272	0.00004	0.05	0.4927	0.0066	12.25	0.18	0.1803	0.0007	2582	29	2655	6	97	
6.1	511	34	0.07	253	0.00002	0.03	0.4811	0.0062	11.80	0.16	0.1779	0.0006	2532	27	2633	6	96	
8.1	220	95	0.43	141	0.00005	0.06	0.5546	0.0084	16.90	0.28	0.2210	0.0011	2844	35	2988	8	95	
8.2	1387	49	0.04	742	0.00012	0.17	0.5237	0.0061	12.90	0.16	0.1787	0.0004	2715	26	2640	4	103	
9.1	188	46	0.24	126	0.00006	0.07	0.5924	0.0116	19.91	0.41	0.2437	0.0012	2999	47	3145	8	95	
9.2	1511	55	0.04	772	0.00002	0.03	0.4994	0.0090	12.37	0.23	0.1796	0.0004	2611	39	2650	4	99	
10.1	640	41	0.06	300	0.00002	0.02	0.4566	0.0064	11.17	0.17	0.1774	0.0009	2425	28	2628	8	92	
11.1	565	39	0.07	261	0.00001	0.01	0.4498	0.0059	10.93	0.16	0.1763	0.0007	2394	26	2618	7	92	
11.2	101	41	0.40	67	0.00015	0.19	0.5774	0.0163	18.38	0.59	0.2308	0.0029	2938	67	3058	20	96	
12.1	631	44	0.07	293	0.00000	0.01	0.4499	0.0082	11.18	0.21	0.1803	0.0006	2395	37	2656	5	90	
13.1	478	29	0.06	228	0.00000	0.00	0.4632	0.0067	11.45	0.18	0.1792	0.0009	2454	30	2646	9	93	
14.1	673	46	0.07	334	0.00002	0.02	0.4817	0.0069	11.91	0.18	0.1794	0.0006	2535	30	2647	6	96	
15.1	804	17	0.02	363	0.00005	0.06	0.4462	0.0063	10.53	0.16	0.1711	0.0008	2379	28	2569	7	93	
16.1	520	23	0.04	247	0.00148	2.04	0.4650	0.0076	11.43	0.22	0.1783	0.0014	2461	34	2637	13	93	

Uncertainties are given at the 1 σ level; *f*_{206%} denotes the percentage of ²⁰⁶Pb that is common Pb; for % Disc, 100 denotes a concordant analysis.

development of S_1 , but to have pre-dated the folding (F_2) of these rocks.

Zircons from sample 9828-189 form a population of very large subhedral grains often exceeding 500 μm in length (Fig. 9b). The Th/U ratios vary from 0.05 to 0.6, with many grains having a Th/U ratio of approximately 0.4 (Table 2). The observed U concentrations are between 200 and 2000 ppm. Zircons from this rock are characterised by a low-CL signature while many zircons lack internal structure (Fig. 9b). Other grains show subtle concentric or sector zoning. In some grains a slightly more luminescent mantle surrounds a lower-CL core, although the opposite relationship is equally common. There is no evidence of the distinct oscillatory zoning observed in the zircons from the enclosing McCue Bluff granite (sample 9828-190).

A total of 11 analyses were carried out for this sample. Eight form a concordant population with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2788 ± 5 Ma (MSWD=0.77; Fig. 9c). The remaining three analyses are slightly discordant and give ages of 3118 ± 5 Ma, 3110 ± 7 Ma and 2616 ± 4 Ma. The older two may have been inherited from the surrounding McCue Bluff granite as a line projected through these data defines an upper intercept near 3150 Ma. The final analysis (2616 ± 4 Ma), when combined with the majority of the data with ages near 2790 Ma, give upper and lower intercepts at 2814 ± 40 Ma and 1600 ± 280 Ma. The high-U and mostly structureless nature of these zircons is characteristic of metamorphic zircon. Thus, the dominant population of zircons with an age of 2788 ± 5 Ma is interpreted to date leucosome formation. Given the structural constraints implied for this sample, it is interpreted that D_{1-2} deformation occurred at c. 2790 Ma.

Sample 9828-218 was taken from a medium-grained biotite-bearing leucosome that crops out at the northern end of McCue Bluff. The leucosome was sampled from a neck of a boudinaged mafic layer (Fig. 9d) found within a D_3 high-strain zone. The leucosome is consequently interpreted to have formed synchronously with D_3 .

Zircons from sample 9828-218 are brown in colour and euhedral in shape. They form a population of large grains averaging 700 μm in length with length/width ratios of approximately 2–3:1 (Fig. 9e). Similar to the previous sample, the CL response of these zircons is low. The zircons from this sample display no internal structure (Fig. 9e), have Th/U ratios typically less than 0.1, and mostly have U concentrations of between 3000 and 6000 ppm (Table 2).

Fifteen analyses from 13 zircon grains were obtained from this sample. All form a single reversely discordant population that yielded a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age

of 2772 ± 4 Ma (MSWD = 7.5; Fig. 9f). By removing the five most scattered analyses, an equivalent $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2775 ± 2 Ma (MSWD = 1.9) is obtained. The first age, obtained using all of the data, is taken to date partial melting and leucosome formation. The age of D_3 deformation is consequently interpreted to have occurred at 2772 ± 4 Ma.

3.3.3. Post-tectonic pegmatite

The final sample from which isotopic data was obtained was a biotite-bearing pegmatite that cropped out at the northern end of the Rimmington Bluff (sample 9828-051). This pegmatite forms an undeformed planar intrusion that cross cuts well-foliated granite gneiss equivalent to sample 9828-039 (3174 ± 9 Ma) described above (Fig. 10a–b). The pegmatite, interpreted to have been emplaced after the D_1 – D_3 stages of deformation, provides a minimum age for these events.

Zircons from sample 9828-051 form a euhedral to subhedral population of pale yellow and translucent grains. These grains show the most distinct core-rim relationships of any of the zircons examined during this study (Fig. 10c). The cores show well-defined oscillatory zoning similar to that seen in the pre-tectonic granite gneiss samples described above. The cores appear broken and/or partially resorbed, and are overgrown by large euhedral low-CL rims. Cores contain between 50 and 200 ppm U and have relatively constant Th/U ratios of about 0.4. In contrast, the rims contain between 500 and 1500 ppm U and have Th/U ratios less than 0.1 (Table 2). The zircons are elongate with length to width ratios of 3:1 and are mostly between 200 and 500 μm long.

Twenty-one analyses of 16 zircon grains were obtained from this sample. Six analyses were of zircon cores, the best four of which yield a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3153 ± 12 Ma (MSWD = 0.94). The remaining two core analyses yielded ages of 2988 ± 8 Ma and 3058 ± 20 Ma, respectively (Fig. 10c). The fifteen rim analyses yielded both concordant and discordant analyses from which a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2645 ± 5 Ma (MSWD = 1.7) was obtained. A discordia fitted to the rim analyses yielded upper and lower intercepts at 2659 ± 12 Ma and 616 ± 260 Ma (Fig. 10c). A similar lower intercept (514 ± 500 Ma) is obtained if a line is fitted to the core data (upper intercept = 3166 ± 100 Ma). These data are interpreted to reflect pegmatite emplacement at 2645 ± 5 Ma based on the age of the high-U and low-luminescent zircon rims. The age and zoning characteristics of the zircon cores suggest the pegmatite was melted from a rock equivalent to the granitic gneiss

Fig. 10. Post-tectonic pegmatite from Rimmington Bluff (sample 9828-051). (a) Outcrop photograph of sample. Geologists circled (bottom centre) for scale. (b) Enlargement of (a). Foliation-parallel amphibolite layers run at a low angle diagonally across the figure. (c) Cathodoluminescent response of representative zircon grains with location and age of individual analyses shown. (d) Wetherall concordia plot of SHRIMP data. Confidence limits for ages and error ellipses are 2σ and 1σ , respectively. White error ellipses are discordant or suspect data that are not used to define weighted mean ages.

into which it intruded. The intrusion of this pegmatite at 2645 ± 5 Ma constrains all stages of pervasive ductile deformation observed along the southern Mawson Escarpment to the Archaean.

4. Discussion of results

The structural and isotopic data presented above suggests that the rocks of the southern Mawson Escarpment underwent a two stage geologic evolution. The earliest event appears to have been magmatic rather than tectonic and to have resulted in the emplacement of the Mawson granitoid suite between 3185 and 3155 Ma. This magmatic event is constrained by the c. 3160 Ma emplacement of the McCue Bluff granite (two samples) and the c. 3175 Ma emplacement of the Rimmington Bluff granite (two samples). A limited component of inherited zircon from sample 9828-190 (McCue Bluff) suggests these intrusions were sourced from slightly older Middle Archaean (c. 3370 Ma) rocks. The intrusion of the Mawson Suite occurred approximately 400 million years before these rocks, and the now interleaved Menzies Series, were deformed and metamorphosed. Given this significant interval, and the lack of control on the depositional age of the Menzies Series, it is permissible that the Mawson granitoids formed an orthogneissic basement onto which the Menzies Series was deposited. This hypothesis has been proposed by a number of earlier studies from the southern Prince Charles Mountains (e.g. Tingey et al., 1981; Tingey, 1982, 1991) and would appear consistent with the detrital ages from the probable correlatives of the Menzies Series exposed at Mt Stinear (Phillips unpublished data). However, there is a distinct lack of evidence, either structurally or in the zircon systematics, for a more protracted history of deformation and metamorphism preserved within the Mawson Suite when compared to that preserved in the interleaved Menzies Series. Furthermore, orthogneissic rocks such as the Mawson Suite require a host into which to intrude. Apart from the Menzies Series, such a sequence of rocks has not been observed or described by this or previous studies. An older sequence of rocks may of course exist, but these are yet to be differentiated from the Menzies Series. At this stage we therefore prefer the interpretation that at least parts of the Menzies Series are at least as old, or older, than the Mawson Suite and that these rocks formed the host into which the Mawson Suite were emplaced.

Irrespective of the exact relationship between the Mawson Suite and the Menzies Series, both sequences of rocks were deformed and metamorphosed at middle to high-grades during a single orogenic event that occurred between 2790 and 2770 Ma. Orogenesis evolved through three distinct phases: (i) low angle shearing and recumbent isoclinal folding (D_{1-2}); (ii) upright folding ($D_{3\text{early}}$) and finally; (iii) strain localisation within steep dipping shear zones ($D_{3\text{late}}$). Over this interval, deformation changed from strongly constrictional with intense linear

fabrics developed during D_{1-2} , to a flattening dominated regime during D_3 . During this latter stage of deformation intense planar fabrics without a strong accompanying lineation developed. D_1 through to D_3 is interpreted to be the result of bulk crustal shortening and SW to S directed tectonic transport (present day coordinates). A similar direction of transport is implied from west to northwest trending recumbent folds described from Mt Menzies (Tingey et al., 1981; Tingey, 1982). Along the southern Mawson Escarpment this event is tightly constrained by the ages obtained from syn- D_2 (2788 ± 5 Ma) and syn- D_3 (2772 ± 4 Ma) leucosomes and more regionally by the widespread resetting of the Rb–Sr system (Tingey et al., 1981; Tingey, 1982, 1991). The emplacement of a post-tectonic pegmatite at 2645 ± 5 Ma confirms that all pervasive ductile deformation along the southern Mawson Escarpment occurred in the Archaean. A similar post-tectonic pegmatite with an age of 2580 Ma (Rb–Sr) places similar constraints on the rocks from Mt Stinear (Tingey, 1991).

Subsequent intrusive activity in the southern Mawson Escarpment is confined to the emplacement of multiple generations of mafic dykes (Fig. 5c). These potentially record one or more periods of Proterozoic crustal extension, although there are no current constraints on how many, or what age these events may have been. The entire block was then dissected along discrete low-grade mylonites (Fig. 5e–f). This deformation may be related to Early Cambrian orogenesis, an event that intensely overprinted the rocks of the Lambert Province to the immediate north (Boger et al., 2001; Boger and Wilson, 2005). However, this is a speculative inference as these structures are yet to be dated. Nevertheless, if these structures do represent an Early Cambrian overprint in the Ruker Complex, it is interesting that evidence for this event is not widely manifested in the zircon systematics. The lower intercepts obtained from sample 9828-051 may point to Early Cambrian lead loss. However, the remaining samples yield mostly Proterozoic lower intercepts. These may reflect the stages of crustal extension implied from the mafic dykes, although realistically the significance, if any, of these lower intercepts is unknown.

5. Wider implications

When one considers the rocks from the southern Mawson Escarpment in the context of the East Antarctic Craton, it is clear that these rocks record a substantially different geologic history when compared to the rocks of the other pre-Gondwana basement Provinces (Fig. 1). The intrusion of the Mawson Suite in the Mid-

dle Archaean (3180–3160 Ma) pre-dates events recorded elsewhere within the East Antarctic Craton. Although metamorphic events of Late Archaean age are known from both the Nimrod Province (c. 2955 Ma; Goodge and Fanning, 1999) and from the Wilkes Province west of the Denman Glacier (c. 2890 Ma; Black et al., 1992a) these events do not temporally correlate with the 2790–2770 Ma orogenic event recorded in the Ruker Complex. Similarly, the widely recognised Palaeoproterozoic events observed in both the Nimrod and Terra Adélie Provinces (1730–1690 Ma) are not recognised in the southern Prince Charles Mountains. Nor for that matter are the Mesoproterozoic events (1340–1290 Ma and 1210–1150 Ma) manifested in the Wilkes Province (Fig. 1). In fact the deposition of the cover sequences in the southern Prince Charles Mountains (the Ruker and Sodruzhestvo Series) quite possibly overlapped temporally with these orogenic events. It is thus clear that the East Antarctic Province is an amalgam of terranes that record quite distinct geologic histories. For such a large craton, this is perhaps not a startling conclusion. The correlation of Proterozoic events from the Nimrod and Terre Adélie Provinces into southern Australia and a similar robust correlation of Mesoproterozoic events from the Wilks Province into southwestern Australia suggest that a large part of the “Mawson Continent” (Fanning et al., 1996) was intact since at least the Middle Proterozoic. The question then becomes when were the distinct Archaean rocks of the Ruker Complex accreted to the remainder of the East Antarctic Craton?

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