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Melt-present shear zones enable intracontinental orogenesis

Sandra Piazolo[1,2*], Nathan R. Daczko[2], David Silva[2] and Tom Raimondo[3]

[1]School of Earth & Environment, University of Leeds, Leeds, LS29JT, UK

[2]Department of Earth & Environmental Sciences, Macquarie University, Sydney, NSW 2109, Australia

[3]School of Natural and Built Environments, University of South Australia, Mawson Lakes, SA 5095, Australia

ABSTRACT

Localized rheological weakening is required to initiate and sustain intracontinental orogenesis, but the reasons for weakening remain debated. The intracontinental Alice Springs Orogen (ASO) dominates the lithospheric architecture of central Australia and involved prolonged (450–300 Ma) but episodic mountain building. The mid-crustal core of the ASO is exposed at its eastern margin, where field relationships and microstructures demonstrate that deformation was accommodated in biotite-rich shear zones. Rheological weakening was caused by localized melt-present deformation coupled with melt-induced reaction softening. This interpretation is supported by the coeval and episodic nature of melt-present deformation, igneous activity and sediment shed from the developing ASO. This study identifies localized melt availability as an important ingredient enabling intracontinental orogenesis.

INTRODUCTION

Collisional mountain belts distal to plate boundaries are commonly referred to as intracontinental orogens (Cunningham, 2005; Aitken, 2011, 2013; Raimondo et al., 2014). Continental plates exhibit prolonged periods of tectonic quiescence at plate interiors, consistent with the scarcity and low magnitude of seismicity, and low maximum horizontal stresses compared to plate boundaries (e.g. Coblentz et al., 1998; Quigley et al., 2010*;* Aitken et al., 2013; Mueller et al., 2012, 2015; Heidbach et al., 2016). This contrasts with strain localization at the 10–100 km scale observed in intracontinental orogens. Whereas lithospheric weakening is essential to enable such strain localization, its root causes remain largely unexplored besides suggestions of gravitational, thermal instabilities and weak/strong provinces (Houseman & Molnar, 2001; Holford et al., 2011, Dyksterhuis & Mueller, 2008).

Central Asia (Tian Shan and Altai) and central Australia (Alice Springs and Petermann Orogens) feature the best known modern and ancient examples of intracontinental orogens, respectively. The Paleozoic Alice Springs Orogen (ASO; Fig. 1) lies within a >1 Ga stable tectonic plate and involved shortening of up to 100 km and deposition of multiple synorogenic, up to 4 km thick sedimentary sequences in line with large scale convergence and mountain building (e.g. Teyssier, 1985; Haines et al., 2001; Klootwijk, 2013; Raimondo et al., 2014). Differential exhumation has resulted in a tilted crustal section exposing the orogenic core; progressively from NW to SE metamorphic grade increases, shear zones widen and their deformation behavior changes from brittle to increasingly ductile (Raimondo et al., 2011; 2014). Whereas upper-crustal processes can be studied in active orogens, the eastern ASO offers the opportunity to investigate the deep-seated mechanics of intracontinental orogenesis. We link field and microstructural observations from a representative high-grade crustal-scale shear zone to episodic orogen-wide igneous activity, localized deformation and metamorphism, and synorogenic sedimentation to evaluate the role of melt availability in enabling intracontinental orogenesis.

SHEAR ZONE CHARACTERISTICS

Regional structure

The ASO is characterized by a pervasive network of NW–SE-trending zones of reverse shear displacement with hydrous mineral assemblages that truncate Paleoproterozoic granulite facies metamorphic fabrics (e.g. Collins & Teyssier, 1989; Cartwright & Buick, 1999; Fig. 1). In the NW, the orogen is ~300 km wide and involves 10-300m wide, reverse-shear zones with < 2 km spacing (Fig. 1c; Collins & Teyssier, 1989). Shear zones are dominated by low-pressure (< 5 kbar) greenschist to lower amphibolite facies mica schists and quartzo-feldspathic mylonites characterized by solid-state deformation and aqueous fluid-rock interaction (e.g., Cartwright & Buick, 1999; Raimondo et al., 2011, 2017). In contrast, the deeper orogenic section to the SE is ~80 km wide, of higher metamorphic grade (6.5–7.0 kbar; 650–700 °C; Mawby et al., 1999; Raimondo et al., 2014) and has a bivergent structure. Strain is localized into 5–8 crustal-scale steep reverse shear zones which are 1–4 km wide, spaced at 8–12 km intervals and dominated by biotite-bearing rocks.

Features of a representative mid-crustal shear zone

The reverse-shear Gough Dam shear zone (GDSZ) is 1–2 km thick, steeply north-dipping (60–80°), E–SE striking (090–150°) over 55 km and juxtaposes two different Proterozoic basement packages (Collins & Shaw, 1995; Fig. 1). It accommodated a dip-slip, reverse-sense displacement of ~40 km based on its steep dip angle, and the number of shear zones and total shortening across the orogen in the SE. Shear strain is

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~20–40 considering its 1–2 km thickness. Shear zone boundaries are abrupt without foliation deflection. Adjacent to these boundaries, anhydrous granulite facies basement rocks show irregularly spaced foliation and cm to dm scale folding with little strain localization along lithological contacts. In contrast, within the shear zone biotite alignment forms a pervasive shear foliation consistent over hundreds of meters across and along strike. Centimeter- to dm-scale compositional banding is foliation-parallel, continuous and varies in biotite mode: three components are distinguished (Fig. 2; see DR1 for details and methods).

Biotite-poor felsic *Component 1* (C1; <5% biotite) includes lenses and layers of varying thickness (0.5–10 cm); they form rootless folds or continuous trains resembling apparent pinch-and-swell structures (Fig. 2a–d); swells and lenses are often connected by mm thick, foliation-parallel biotite-rich seams. C1 is K-feldspar- and quartz-rich and frequently bordered by biotite-rich selvedges (Fig. 2d–e; DR2). Feldspar grains have a small grain size range (2–3 mm), may be rectangular, interlocking and without clear crystallographic or shape preferred orientation (DR2). They occur with interstitial quartz and less commonly plagioclase. Interstitial grains show connectivity in 3D (i.e. interstitial grains that are spatially separate in 2D sections exhibit the same crystallographic orientation), often aspect ratios > 8 and low dihedral angles (<60°, Fig. 2e). Irregular boundaries may also occur (Fig. 2e).

Finer-grained (~1 mm) granitic *Component 2* (C2; 5–25% biotite) constitutes most of the shear zone (Fig. 2), forms foliation-parallel dm thick continuous bands and displays the same microstructures observed in C1 along with "string of beads" textures (i.e. an array of quartz grains along grain boundaries; Fig. 2f).

Biotite-rich *Component 3* (C3, >50 %biotite) is seen as selvedges around C1 and as mm-thick seams and continuous cm- to m-thick glimmerite bands in C2 (Fig. 2a–d).

Biotite is medium grained (1 mm), aligned foliation-parallel and rarely kinked and bent (Fig. 2g). Quartz is interstitial with aspect ratios >8 (Fig. 2g) and rectangular single K-feldspar grains and clusters of interlocked feldspar grains are seen in the glimmerite matrix (Fig. 2c). The glimmerite contains 1–3 additional minor phases e.g. interstitial muscovite, elongate sillimanite clusters (Fig. 2g).

TEMPORAL PATTERNS OF OROGENIC ACTIVITY

A compilation of orogen-wide geochronological datasets shown in Fig. 3 and DR3 demonstrates an episodic temporal evolution of the ASO. The data exhibit overlap of igneous activity (*Ig*), derived from peraluminous granitic pegmatite dykes (Buick et al., 2008) and decameter scale granitic plugs (Buick et al., 2001), with metamorphic and deformation ages of shear zone rocks (*D*). Geochronological data shows that some shear zones are active during separate episodes of orogenic activity (e.g. Raimondo et al., 2014; DR3). The timing of peaks in synorogenic sedimentation (*S*) at c. 450–435 Ma, c. 385–365 Ma and c. 340–315 Ma coincides with three of the igneous and tectonometamorphic episodes.

DISCUSSION

Melt-present deformation

The GDSZ lacks microstructural evidence of solid-state, crystal-plastic deformation, e.g. mantled porphyroclasts, bimodal grain size distribution, undulose extinction or kinking (Fig. 2, DR1, Passchier and Trouw 2005). Instead, microstructures indicative of the former presence of melt are preserved. These include disequilibrium grain boundaries (dashed lines; Fig. 2e–f), interstitial minerals with low dihedral angles, grains with aspect ratios > 8 and “string of beads” textures; the latter two textures are interpreted as remnants of former grain boundary melt films (Vernon, 2011; Holness et al. 2011). Component C1 also exhibits typical igneous features including interstitial quartz and interlocking, rectangular feldspar crystals with unimodal grain size and no clear preferred orientation (inset Fig. 2d, DR2).

Recent work has shown that during melt-present deformation, microstructures indicative of the former presence of melt are formed and preserved (Stuart et al., 2018a,b; Meek et al., 2019). In such shear zones, strain is dominantly accommodated by viscous melt flow, resulting in a viscosity drop of at least one order of magnitude relative to solid-state deformation (Lee et al., 2018; Pakrash et al., 2018). If annealing had been an important process, subtle microstructures such as interstitial grains with low dihedral angles and “string of beads” textures would likely have been erased (Piazolo et al. 2006).

The ubiquitous evidence of former melt presence within the GDSZ focuses attention on the origin of the melt. The anhydrous and infertile nature of the granulite facies basement rocks that host the shear zone is inconsistent with local melt derivation (Buick et al., 2008) and strongly contrasts with the abundance of hydrous minerals in the deformed rocks. Biotite growth during shearing and a lack of partial melting hallmarks (e.g., presence of peritectic minerals and reaction textures indicating the consumption of hydrous phases) further argue against in-situ melting of the host rocks. This suggest that melt was externally derived, implying syntectonic melt flux through the shear zone.

Melt-present shearing and melt–rock interaction facilitate enhanced rheological weakening

In the GDSZ biotite-rich selvedges and felsic igneous lenses are spatially linked (Fig. 2d), individual lenses are commonly connected by biotite seams and cm- to dm-thick biotite-rich glimmerite bands invariably contain some igneous features (Fig. 2c,g). This spatial association suggests a causal relationship, whereby biotite formed due to interaction between a hydrous melt and the host rock. Like a dynamic version of the hydration crystallization reactions of Beard et al. (2004), disequilibrium during melt-rock interaction drives the dissolution of precursor granulite and the precipitation of a biotite-rich assemblage in equilibrium with the fluxing melt (Stuart et al., 2016, 2017; Meek et al., 2019). This results in melt-mediated replacement reactions (Daczko et al., 2016), analogous to lower temperature aqueous fluid-mediated replacement reactions (Putnis, 2009).

Hence, in the biotite-rich glimmerite bands, melt flux drove widespread melt–rock interaction and metasomatism, producing a locally hydrated rock. Experiments show that ascending peraluminous melts become increasingly reactive (Clemens, 2003) explaining widespread melt-induced reactions during melt flux. Exothermic biotite growth (Haack & Zimmerman, 1996) and melt flux-related heating maintained sufficiently high temperatures to limit melt crystallization during reaction and ascent. Deformation inconsistencies between adjacent grains caused increased porosity and permeability and enhanced melt migration (Menegon et al. 2015). Additionally, deformation assisted melt expulsion by synkinematic filter pressing (Park & Means, 1996), explaining the observed low abundance of preserved quartzo-feldspathic igneous material in the glimmerite bands.

Development of the biotite-rich glimmerite bands represents a form of reaction softening (Watts & Williams, 1983), commonly attributed to aqueous fluid–rock interaction (Teall, 1885; Brodie & Rutter, 1985). We argue that melt–rock interaction is the root cause of mica growth and drives concomitant rheological weakening. Fluid (including melt) cannot support shear stresses; hence, melt-bearing rocks are intrinsically weak (e.g., van der Molen & Paterson, 1979; Rutter & Neumann, 1995). Rosenberg & Handy (2005) show that >7 vol.% melt weakens rocks by at least one order of magnitude. This observation has been extended to shear zones to infer rheological weakening of 1–2 orders of magnitude at the km-scale (Brown & Solar, 1998; Marchildon & Brown, 2003; Weinberg & Mark, 2008; Jamieson et al., 2011) and orogen-scale (Hollister & Crawford, 1986; Harris, 2007). Fluid overpressure (Hubbert & Rubey, 1959) and melt-induced heating may enhance weakening within a deforming rock (Tommasi et al., 1994). We therefore interpret that the ASO shear zones were weak during episodes of melt flux due to melt-present deformation and melt-induced reaction softening.

Orogenic episodicity: dynamic feedback between far-field stresses, melt availability and rheological weakening

The intracontinental setting of the ASO suggests that orogenesis involved a localized weak zone within otherwise strong lithosphere. Recent work shows that the failed Larapinta Rift (LR; Fig. 3), together with its thick sedimentary fill, represents the required weak zone and, additionally, global-scale plate reorganization drove rift-inversion and strain localization (Silva et al., 2018). Whereas the importance of fault reactivation in controlling deformation patterns during orogenesis has been recognized (e.g. De Graciansky et al. 2010, Kober et al. 2013), the episodic nature of the ASO (Fig. 3) suggests another factor must play a role. If orogenesis was enabled solely by reactivation of pre-existing faults, strain localization at rheological interfaces or in the presence of weak rift fill sediments, then continuous deformation may be expected. In contrast, magmatic systems are inherently episodic due to the link between melt pressure build-up, release and melt extraction (e.g. Schmeling 2006).

Episodic melt-present deformation requires a fertile source, a mechanism to cause melting, and alternating melt-present weakening/deformation and fault strengthening. For the ASO, a fertile source has been contentious, as hydrous shear zones cut anhydrous granulite (Buick et al., 2008); proposed candidates include unmetamorphosed sedimentary rocks of the neighboring intracratonic Neoproterozoic–Paleozoic Amadeus and Georgina Basins (Fig. 1), or the Cambrian–Ordovician Harts Range Group deposited in the epicratonic Larapinta rift basin (Irindina Province, Fig. 1; Maidment et al., 2013). These sources require either tectonic underthrusting (Buick et al., 2008) or deep burial (Tucker et al., 2015) to supply fertile crust to deeper structural levels and facilitate melt production.

We argue that in the deep crust, recently deposited sediments underwent prograde metamorphism to produce melt over a sustained period, but that melt migration and associated deformation was only facilitated when (a) an external force was applied due to plate reorganization and associated continuous external stresses; and (b) melt pressure was sufficiently high to overcome yield stresses. During melt release, deformation was highly localized and melt-present (see experiments by Zanella et al., 2015), facilitating fault reactivation in zones of relative weakness (biotite-rich shear zones rather than strong granulites). Upon melt source drainage (Schmeling 2006), shear zones became strong and deformation ceased, allowing melt pressure to gradually increase. Conceptually this is similar to the fault valve fluid-deformation cycles envisaged by Sibson (1990). Once melt pressure was sufficiently high, shear zones became active again, initiating another episode of simultaneous igneous and tectonic activity. By this mechanism, repeated episodes of melt-present deformation and melt-induced reaction softening drove substantial weakening, and thus melt availability played a critical role in enabling intracontinental orogenesis.

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REFERENCES

Aitken, A. R. A., 2011, Did the growth of Tibetan topography control the locus and evolution of Tian Shan mountain building?: Geology, v. 39, p. 459–462.

Aitken, A. R. A., Raimondo, T., & Capitanio, F. A. 2013, The intraplate character of supercontinent tectonics: Gondwana Research, v. 24, p. 807–814.

Beard, J. S., Ragland, P. C., & Rushmer, T., 2004, Hydration crystallization reactions between anhydrous minerals and hydrous melt to yield amphibole and biotite in igneous rocks: description and implications: The Journal of Geology, v. 112, p. 617–621.

Brodie, K. H., & Rutter E. H., 1985, On the relationship between deformation and metamorphism, with special reference to the behaviour of basic rocks. In:

Metamorphic Reactions: Kinetics, Textures and Deformation (eds Thompson, A. B. and Rubie, D. C.): Advances in Physical Geochemistry, v. 4, p. 138–179.

Brown, M., & Solar, G. S., 1998, Shear-zone systems and melts: feedback relations and self-organization in orogenic belts: Journal of Structural Geology, v. 20, p. 211-227.

Buick, I. S., Miller, J. A., Williams, I. S., & Cartwright, I., 2001, Ordovician high-grade metamorphism of a newly recognised late Neoproterozoic terrane in the northern Harts Range, central Australia. Journal of Metamorphic Geology, v. 19, p. 373–394.

Buick, I. S., Storkey, A., & Williams, I. S., 2008, Timing relationships between pegmatite emplacement, metamorphism and deformation during the intra-plate Alice Springs Orogeny, central Australia: Journal of Metamorphic Geology, v. 26, p. 915–936.

Cartwright, I., & Buick, I. S., 1999, Meteoric fluid flow within Alice Springs age shear zones, Reynolds Range, central Australia: Journal of Metamorphic Geology, v. 17, p. 397–414.

Clemens, J. D., 2003, S-type granitic magmas—petrogenetic issues, models and evidence: Earth-Science Reviews, v. 61, p. 1-18.

Coblentz, D. D., Zhou, S., Hillis, R. R., Richardson, R. M., Sandiford, M., 1998, Topography, boundary forces, and the Indo-Australian intraplate stress field: Journal of Geophysical Research, Solid Earth, v. 103, p. 919-931.

Collins, W. & Shaw, R., 1995, Geochronological constraints on orogenic events in the Arunta inlier: A review: Precambrian Research, v. 71, p.315–346.

Collins, W. & Teyssier, C., 1989. Crustal scale ductile fault systems in the Arunta inlier, central Australia: Tectonophysics, v. 158, p. 49–66.

Collins, W. J., & Teyssier, C., 1989, Crustal scale ductile fault systems in the Arunta Inlier, central Australia: Tectonophysics, v. 158, p. 49-66.

Cunningham, D., 2005, Active intracontinental transpressional mountain building in the Mongolian Altai: Defining a new class of orogen: Earth and Planetary Science Letters, v. 240, p. 436–444.

Daczko, N. R., Piazolo, S., Meek, U., Stuart, C. A., & Elliott, V., 2016, Hornblendite delineates zones of mass transfer through the lower crust: Scientific reports, 6.

De Graciansky, P. C., Roberts, D. G., & Tricart, P., 2010. The late Cretaceous phase and Onset of Alpine shortening. In: "The Western Alps, from rift to passive margin to orogenic belt: an integrated geoscience overview": v. 14, p. 169-182.

Dyksterhuis, S., & Müller, R. D., 2008, Cause and evolution of intraplate orogeny in Australia: Geology, v. 36, p. 495–498.

Haack, U. K., & Zimmermann, H. D., 1996, Retrograde mineral reactions: a heat source in the continental crust?: Geologische Rundschau, v. 85, p. 130–137.

Haines, P. W., Hand, M., & Sandiford, M., 2001, Palaeozoic synorogenic sedimentation in central and northern Australia; a review of distribution and timing with implications for the evolution of intracontinental orogens: Australian Journal of Earth Sciences, v. 48, p. 911–928.

Harris, N., 2007, Channel flow and the Himalayan-Tibetan orogen: A critical review: Journal of the Geological Society, v. 164, p. 511–523.

Heidbach, O., Rajabi, M., Reiter, K., & Ziegler, M., 2016, World stress map 2016: Science, v. 277, p. 956-1.

Hollister, L. S., & Crawford, M. L., 1986, Melt-enhanced deformation: A major tectonic process: Geology, v. 14, p. 558–561.

Holness, M. B., Cesare, B., & Sawyer, E. W., 2011, Melted rocks under the microscope: Microstructures and their interpretation: Elements, v. 7, p.247- 252.

Holford, S. P., Hillis, R. R., Hand, M. & Sandiford, M., 2011. Thermal weakening localizes intraplate deformation along the southern Australian continental margin. Earth and Planetary Science Letters, v. 305, p. 207-214.

Houseman, G., & Molnar, P., 2001, Mechanisms of lithospheric rejuvenation associated with continental orogeny: Geological Society, London, Special Publications, v. 184, p. 13–38.

Hubbert, M. K., & Rubey, W. W., 1959, Role of pore fluid pressure in the mechanics of overthrust faulting, *in* Mechanics of fluid-filled porous solids and its application to overthrust faulting: Geological Society of America Bulletin, v. 70, p. 115–205.

Jamieson, R.A., Unsworth, M.J., Harris, N.B., Rosenberg, C.L. & Schulmann, K., 2011, Crustal melting and the flow of mountains: Elements, v. 7, p. 253–260.

Klootwijk, C., 2013, Middle–Late Paleozoic Australia–Asia convergence and tectonic extrusion of Australia: Gondwana Research, v. 24, p. 5–54.

Kober, M., Seib, N., Kley, J., & Voigt, T., 2013. Thick-skinned thrusting in the northern Tian Shan foreland, Kazakhstan: structural inheritance and polyphase deformation. Geological Society, London, Special Publications, v. 377, p. 19-42

Lee, A. L., Torvela, T., Lloyd, G. E., & Walker, A. M. (2018). Melt organisation and strain partitioning in the lower crust: Journal of Structural Geology, v. 113, p. 188-199.

Maidment, D. W., Hand, M. & Williams, I. S., 2013. High grade metamorphism of sedimentary rocks during Palaeozoic rift basin formation in central Australia: Gondwana Research, v. 24, 865-885.

Marchildon, N., & Brown, M., 2003, Spatial distribution of melt-bearing structures in anatectic rocks from Southern Brittany, France: implications for melt transfer at grain-to orogen-scale: Tectonophysics, v. 364, p. 215–235.

Mawby, J., Hand, M. & Foden, J. 1999. Sm–Nd evidence for high-grade Ordovician metamorphism in the Arunta Block, central Australia: Journal of Metamorphic Geology, v. 17.

Meek, U., Piazolo, S., & Daczko, N. R., 2019. The field and microstructural signatures of deformation-assisted melt transfer: insights from magmatic arc lower crust, New Zealand: Journal of Metamorphic Geology, v. 37, p. 795-821.

Menegon, L., Fusseis, F., Stünitz, H., & Xiao, X., 2015. Creep cavitation bands control porosity and fluid flow in lower crustal shear zones: Geology, v.43, p. 227-230.

Mueller, R., Dyksterhuis, S., Rey, P., 2012, Australian paleo-stress fields and tectonic reactivation over the past 100 Ma: Australian Journal of Earth Sciences, v. 59, p. 13-28.

Mueller, R., Yatheesh, V., Shuhail, M., 2015. The tectonic stress field of India since the Oligocene: Gondwana Research, v. 28, p. 612-624.

Park, Y., & Means, W. D., 1996. Direct observation of deformation processes in crystal mushes: Journal of Structural Geology, v. 18, p. 847–858.

Passchier, C. W., & Trouw, R. A., 2005. Microtectonics. Springer Science & Business Media.

Piazolo, S., Bestmann, M., Prior, D.J. and Spiers, C.J., 2006. Temperature dependent grain boundary migration in deformed-then-annealed material: observations from experimentally deformed synthetic rocksalt: Tectonophysics, v. 427, p. 55-71.

Prakash, A., Piazolo, S., Saha, L., Bhattacharya, A., Pal, D. K., & Sarkar, S., 2018. Deformation behavior of migmatites: insights from microstructural analysis of a garnet–sillimanite–mullite–quartz–feldspar-bearing anatectic migmatite at Rampura–Agucha, Aravalli–Delhi Fold Belt, NW India: International Journal of Earth Sciences, v. 107, p. 2265-2292.

Putnis, A., 2009. Mineral replacement reactions: Reviews in mineralogy and geochemistry, v. 70, p. 87–124.

Quigley, M.C., Clark, D., Sandiford, M., 2010. Tectonic geomorphology of Australia: Geological Society, London, Special Publications, v. 346, p. 243–265

Raimondo, T., Clark, C., Hand, M., & Faure, K., 2011, Assessing the geochemical and tectonic impacts of fluid–rock interaction in mid-crustal shear zones: a case study from the intracontinental Alice Springs Orogen, central Australia: Journal of Metamorphic Geology, v. 29, p. 821–850.

Raimondo, T., Hand, M., & Collins, W. J., 2014, Compressional intracontinental orogens: Ancient and modern perspectives: Earth-Science Reviews, v. 130, p. 128–153.

Raimondo, T., Payne, J., Wade, B., Lanari, P., Clark, C., & Hand, M., 2017, Trace element mapping by LA-ICP-MS: assessing geochemical mobility in garnet: Contributions to Mineralogy and Petrology, v. 172, p. 17-38.

Rosenberg, C. L., & Handy, M. R., 2005, Experimental deformation of partially melted granite revisited: implications for the continental crust: Journal of Metamorphic Geology, v. 23, p. 19–28.

Rutter, E. H., & Neumann, D. H. K., 1995, Experimental deformation of partially molten Westerly granite under fluid-absent conditions, with implications for the extraction of granitic magmas: Journal of Geophysical Research: Solid Earth, v. 100, p. 15697-15715.

Shaw, R. D., Zeitler, P. K., McDougall, I., & Tingate, P. R., 1992, The Palaeozoic history of an unusual intracratonic thrust belt in central Australia based on 40Ar-39Ar, K-Ar and fission track dating: Journal of the Geological Society, v. 149, p. 937-954.

Sibson, R. H., 1990. Conditions for fault-valve behavior: Geological Society, London, Special Publications, v. 54, p. 15-28.

Silva, D., Piazolo, S., Daczko, N. R., Houseman, G., Raimondo, T., & Evans, L., 2018, Intracontinental Orogeny Enhanced by Far-Field Extension and Local Weak Crust: Tectonics, v. 37, p. 4421-4443.

Stuart, C. A., Piazolo, S., & Daczko, N. R., 2016, Mass transfer in the lower crust: Evidence for incipient melt assisted flow along grain boundaries in the deep arc granulites of Fiordland, New Zealand: Geochemistry, Geophysics, Geosystems, v. 17, p. 3733–3753.

Stuart, C. A., Daczko, N. R., & Piazolo, S., 2017, Local partial melting of the lower crust triggered by hydration through melt–rock interaction: an example from Fiordland, New Zealand: Journal of Metamorphic Geology, v. 35, p. 213–230.

Stuart, C. A., Meek, U., Daczko, N. R., Piazolo, S., & Huang, J. X., 2018a, Chemical signatures of melt–rock interaction in the root of a magmatic arc: Journal of Petrology, v. 59, p. 321–340.

Stuart, C. A., Piazolo, S., & Daczko, N. R., 2018b, The recognition of former melt flux through high-strain zones: Journal of Metamorphic Geology, v. 36, p. 1049-1069.

Teall, J. J. H., 1885, The metamorphism of dolerite into hornblende schist: Quart. J. Geol. Soc. London, *v.*41, p. 133–145.

Teyssier, C., 1985. A crustal thrust system in an intracratonic tectonic environment: Journal of Structural Geology, v. 7, p. 689–700.

Tommasi, A., Vauchez, A., Femandes, L. A., & Porcher, C. C., 1994, Magma-assisted strain localization in an orogen-parallel transcurrent shear zone of southern Brazil: Tectonics, v. 13, p. 421-437.

Tucker, N. M., Hand, M., & Payne, J. L., 2015, A rift-related origin for regional medium-pressure, high-temperature metamorphism: Earth and Planetary Science Letters, v. 421, p. 75–88.

Van der Molen, I., & Paterson, M. S., 1979, Experimental deformation of partially-melted granite: Contributions to Mineralogy and Petrology, v. 70, p. 299–318.

Vernon, R. H., 2011, Microstructures of melt-bearing regional metamorphic rocks: Geological Society of America Memoirs, v. 207, p. 1–11.

Watts, M.J., & Williams, G.D., 1983, Strain geometry, micro- structure and mineral chemistry in metagabbro shear zones: a study of softening mechanisms during progressive mylonitization: Journal of Structural Geology, v. 5, p. 507–517.

Weinberg, R. F., & Mark, G., 2008, Magma migration, folding, and disaggregation of migmatites in the Karakoram Shear Zone, Ladakh, NW India: Geological Society of America Bulletin, v. 120, p. 994-1009.

Zanella, A., Cobbold, P. R., & de Veslud, C. L. C., 2014, Physical modelling of chemical compaction, overpressure development, hydraulic fracturing and thrust detachments in organic-rich source rock: Marine and Petroleum Geology, v. 55, p.262-274.

FIGURE CAPTIONS

Figure 1. Geological context of the Alice Springs Orogen (ASO). (a,b) Lateral extent and position of the orogen within the Paleozoic Australian continent (white outline with major terranes shown); note the ASO was approximately 1000 km inland from the W, N and S continental margins and 200–300 km from the E. The intracratonic sedimentary basins of central Australia indicate the footprint of the former Centralian Superbasin and key ASO synorogenic depocenters; (c) Generalised geological map of the eastern Arunta Region; note the Gough Dam shear zone field site indicated with a gold star (23.147 °S, 134.565 °E); (d) Cross-section along profile X–Y indicated in (c), showing a bivergent crustal-scale pop-up structure. Figure modified from Collins & Teyssier (1989), Raimondo et al. (2011) and Scrimgeour (2013). AB–Amadeus Basin, AR–Arunta Region, CB–Canning Basin, EGC–Entia Gneiss Complex, GB–Georgina Basin, GDSZ–Gough Dam shear zone, HRMC–Harts Range Metaigneous Complex, MP–Musgrave Province, NB–Ngalia Basin, OB–Officer Basin, RR-AR–Reynolds-Anmatjira Ranges, SMC–Strangways Metamorphic Complex, SR–Strangways Range, WB–Wiso Basin.

Figure 2. Field and microstructural relationships of the Gough Dam shear zone; S1 represents the shear foliation; see DR1 for further details. (a) Outcrop-scale features showing fine-grained, well-foliated granitic bands with biotite-rich glimmerite (white arrows) and trains of biotite-poor components (black arrows); scale bar is 1 m; (b,c) Fine-grained and well-foliated granitic component with elongate lens-shaped biotite-

poor component (rare rootless folds) and biotite-rich glimmerite components (white arrows). (d) Outcrop-scale and detailed views of the granitic component with a high abundance of lens-shaped biotite-poor felsic components with pronounced biotite-rich selvedges; (e–f) Photomicrographs and detailed views featuring K-feldspar–quartz-rich lens (e) and granitic component (f) showing interlocked rectangular K-feldspar grains with quartz grain boundary films and an array of 0.1 - 0.2 mm, equidimensional quartz grains decorating grain boundaries forming "string of beads" textures. Inset to right shows interstitial K-feldspar with low apparent dihedral angles (LDA) between "fingers" of a single quartz grain (LDA q-k-q); all minerals display limited evidence of crystal-plastic deformation (i.e. undulose extinction, subgrain boundaries, core-mantle structures, shape and crystallographic preferred orientation etc.) and have irregular disequilibrium boundaries (dashed lines); (g) Photomicrographs of biotite-rich glimmerite components showing assemblages including biotite, muscovite, quartz and sillimanite; quartz and biotite are intergrown such that quartz displays low apparent dihedral angles between two biotite grains (LDA b-q-b); all minerals display limited evidence of crystal-plastic deformation (i.e. grains lack undulose extinction, kinking, etc.).

Figure 3. Gaussian-summation probability density distribution plots of orogen-wide age data for synorogenic sedimentation (*S*, orange), deformation/metamorphism (*D*, blue) and igneous activity (*Ig*, red) (for methods and tabulated data see DR3). Synorogenic sedimentation intervals are based on biostratigraphic evidence in orogenic deposits (Haines et al., 2001; Shaw et al., 1992), metamorphic and deformation ages are established by syntectonic monazite, garnet and mica geochronology (DR3 and related references), and igneous activity is derived from

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peraluminous granitic pegmatite dykes (Buick et al., 2008) and minor plutons (Buick et al., 2001). North-south extensional activity formed the deep epicratonic Larapinta Rift (LR; 460–480 Ma) and immediately preceded the compressional Alice Springs Orogeny. The timing of orogenic activity along the eastern margin of Australia in the Tasmanides (Tasman.) is dominantly extensional, with periods of compression (marked as “x”) or significant strike slip movement (double arrows) shown (Raimondo et al., 2014; Silva et al., 2018).

[1]GSA Data Repository item 201xxxx, Appendix containing Table DR1 (characteristics of compositionally and texturally distinct rock components identified in outcrop and petrographic thin sections), Figure DR2 (petrographic image) and Table DR3 (age compilation table including methods), is available online at www.geosociety.org/pubs/, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.