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Precambrian faulting episodes and insights into the tectonothermal history of North Australia: Microstructural evidence and K–Ar, ^{40}Ar–^{39}Ar, and Rb–Sr dating of syntectonic illite from the intracratonic Millungera Basin.

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ABSTRACT

Australian terranes concealed beneath Mesozoic cover record complex Precambrian tectonic histories involving a successive development of several Proterozoic to Paleozoic orogenic systems. This study presents an integrated approach combining K–Ar, ^{40}Ar–^{39}Ar, and Rb–Sr geochronology of Precambrian authigenic illites from the recently discovered Millungera Basin in north-central Australia. Brittle deformation and repeated fault activity are evident from the sampled cores and their microstructures, probably associated with the large-scale faults inferred from interpretations of seismic survey. Rb–Sr isochron, ^{40}Ar–^{39}Ar total gas, and K–Ar ages are largely consistent indicating late Mesoproterozoic and early Proterozoic episodes (~1115 ± 26 Ma, ~1070 ± 25 Ma, ~1040 ± 24 Ma, ~1000 ± 23 Ma, and ~905 ± 21 Ma) of active tectonics in north-central Australia. K–Ar results show that illites from fault gouges and authigenic matrix illites in undeformed adjacent sandstones precipitated contemporaneously, indicating that advection of tectonically mobilised fluids extended into the undeformed wall rocks above or below the fracture and shear (fault gouge) zones. This study provides insight into the enigmatic time-space distribution of Precambrian

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tectonic zones in central Australia, which are responsible for the formation of a number of sedimentary basins with significant energy and mineral resources.

Keywords: Fault gouge; illite; micro structure; fluid flow; isotope dating; Precambrian; Australia.

1. INTRODUCTION

Direct dating of brittle faulting is crucial for determining the absolute timing of inscrutable time-space distribution of tectonothermal events in concealed Precambrian terranes. Over the last decade, dating of illitic clay from near-surface fault gouges has increasingly become a routine approach to defining the timing of brittle deformations (van der Pluijm et al., 2001; Uysal et al., 2006; Mutlu et al., 2010; Zwingmann and Mancktelow, 2004, Zwingmann et al., 2010; Duvall et al., 2011; Hetzel et al., 2013, Torgersen et al., 2014; Mancktelow et al., 2016, Viola et al., 2016; Babaahmadi et al., 2019). This technique has been particularly useful in better understanding the development of convergent plate boundaries and continental collisions (e.g., van der Pluijm et al., 2001; Duvall et al., 2011; Isik et al., 2015; Babaahmadi et al. 2019), movements along transform plate margins (Uysal et al., 2006; Mutlu et al., 2009; Boles et al., 2015), and the formation of orocline bending accompanied by regional strike-slip faulting (Rosenbaum et al., 2015).

While fault gouges reported by earlier studies were mainly from surface outcrops, dating of concealed fault systems is more challenging due to the lack of direct structural observations. Although unknown fault systems buried under thick sedimentary basins can be denoted by geophysical techniques such 2D and 3D seismic reflections, cores from boreholes shaft or tunnel sites intersecting fault zones can be used to date fault reactivation episodes (e.g., Viola et al., 2013; Yamasaki et al., 2013, Elminen et al., 2018). The current study investigates fault rocks and the host sandstone intersected in drill cores from the newly

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discovered Millungera Basin in north Queensland, north-central Australia (Fig. 1). It demonstrates how illite geochronology in combination with microstructural and mineralogical studies can be used to define a concealed, previously unrecorded Proterozoic tectonic events. Prior to this study (Fig. 1), almost no geological information was available on the Precambrian geology of large parts of north-central Australia, including the Millungera Basin, except for some regional geophysical data (Korsch et al., 2011, 2012) (Fig. 1). This is due to an extensive cover of sediments of the Jurassic–Cretaceous Eromanga-Carpentaria Basin (Fig. 1). Further uncertainties in tectonic interpretation of Australian Precambrian terranes arises from the tendency for original tectonic information to be masked by younger tectonics. Therefore, a major objective of this study was to provide insight into the enigmatic time-space distribution of Mid- to Late- Mesoproterozoic tectonic zones in central Australia, which are responsible for the formation of a number of sedimentary basins with significant potential for energy and mineral resources (Korsch et al., 2011, 2012).

Many previous studies have largely focussed on shallow crustal faults that form at diagenetic temperatures below 200°C. Fault gouges from such environments are assumed to consist of (1) detrital illite/muscovite ($2M_1$) derived from wall rocks and (2) authigenic or in situ illite ($1M/M_d$) precipitated within the brittle fault zone during faulting (van der Pluijm et al., 2001; Duvall et al., 2011). Based on a two-end member mixing model, quantified percentages of each illite polytypes (1M and 2M) in different clay size fractions and their apparent $^{40}Ar–^{39}Ar$ ages are used to extrapolate the age of the pure authigenic clay (IAA: Illite Age Analyis approach, e.g., van der Pluijm et al., 2001; Duvall et al., 2011). However, assuming that 2M illite is systematically of detrital origin can be misleading, since the formation of authigenic 2M illite in diagenetic-to-hydrothermal conditions is also reported in the literature (e.g, Lonker and Gerald, 1990; Clauer and Liewig, 2013) and brittle faulting can produce authigenic 2M illite particularly in areas of elevated geothermal gradients or deeper

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parts of exhumed faults (Zwingmann et al., 2010; Viola et al., 2013; Mancktelow et al., 2015). While successful isotopic dating of brittle faulting and reactivation within single fault core was reported previously (Viola et al., 2013), the present study integrating fault rocks from different depths and locations is a new and challenging approach to help better understand illite crystallisation in gouge during relatively low-temperature brittle fault reactivation episodes in complex Precambrian tectonic settings.

2. GEOLOGICAL SETTING, SAMPLE LOCATIONS AND SAMPLING

The Millungera Basin is a recently discovered sedimentary basin in north Queensland, Australia (Korsch et al., 2011, Fig. 1). It occurs to the east of the Paleoproterozoic Mount Isa Province and is covered by the thin Jurassic–Cretaceous Eromanga-Carpentaria Basin. An angular unconformity between the Eromanga and Millungera Basins indicates that the upper part of the Millungera Basin was eroded prior to deposition of the Eromanga-Carpentaria Basin (Korsch et al., 2011), allowing sampling of the deeper part of the basin, which is strongly deformed and faulted (see below). Interpretation of aeromagnetic data suggests that the basin might have a ~~real~~ dimension of up to 280 km by 95 km (Korsch et al., 2011). Apart from geophysical data, almost no geological information exists on the basin. Prominent thrust fault systems truncate both the western and eastern margins of the basin. Particularly the eastern part of the basin has been cut by several deep-penetrating, northeast-dipping thrust faults, with associated development of hanging wall anticlines. Based on SHRIMP U-Pb geochronology of detrital zircons from the Millungera Basin sandstones, the maximum depositional age of the Millungera Basin is constrained to 1574 ± 14 Ma (Neumann and Kositcin, 2011).

Core samples were taken from the lower parts of boreholes Julia Creek 1 (JC) and Dobbyn 2 (Dob) drilled as part of a state (Queensland)-wide geothermal investigation. The

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wells are 150 km apart (Fig. 1 and Table 1) and intersect the Mesozoic Eromanga-Carpentaria Basin in the upper part. Julia Creek 1 intersected 320.05m of the Eromanga Basin sequence and 179.97m of the Millungera Basin sequence; Dobbyn 2 intersected 332.40m of the Carpentaria Basin sequence and 155.64m of the Millungera Basin sequence. It should be noted that the succession within the Millungera Basin has not been formally defined. According to deep seismic reflection survey, a number of large-scale structures are interpreted to occur as basin-bonding and intra-basin fault systems (see Korsch et al., 2011). Small scale faults and fractures have also been described from logging of the cores extracted from JC and Dob (Faulkner et al., 2012; Fitzell et al., 2012). We collected a total of 9 Julia Creek 1 and 6 Dobbyn 2 fault gouge samples that were all analysed for the <2 μm clay mineral content (Table 1; Supplementary Fig. S1) and some of which have been selected for K–Ar, ^{40}Ar–^{39}Ar, and Rb–Sr dating and trace element studies. We also sampled representative host rock samples adjacent to the fault gouge zones (Table 1).

3. ANALYTICAL PROCEDURES

3.1. Clay characterisation

Samples were prepared for clay-fraction separation by gently hand-crushing the rocks to sand size to avoid artificially reducing grainsize of detrital/primary mineral components. Samples were then disaggregated in distilled water using an ultrasonic bath. Clay fractions were separated by the sedimentation method (for <2 μm) and centrifugation (for <1 μm subfractions to <0.1 μm). XRD on whole-rock samples and clay separates of different size fractions were carried out (Table 1). The XRD analyses were conducted on a Bruker D4 Endeavor and D8 Advance (CoKα and CuKα radiation, respectively), operated at 40 kV and 30 mA at a scanning rate of 1°2θ/min and 0.05°/step. Following XRD analysis of air-dried samples, the oriented clay-aggregate mounts were placed in an ethylene–glycol atmosphere at

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30–40°C overnight prior to additional XRD analyses. For polytype analyses, clay fractions of random powder from fault gouge samples (if sufficient amount of material was available) were scanned from 16 to 44 °2θ in the step-scanning mode with a step size of 0.05 degrees and a counting time of 30 second per step.

Illite polytypes for randomly oriented pure illite samples have been distinguished with the diagnostic peaks suggested by Grathoff and Moore (1996). To determine the $2M_1$ and $1M$ and $1M_d$ % contents of illite/muscovites, the ratios of (2.80 Å – 3.0 Å)/(2.58 Å) and (3.07 Å)/(2.58 Å) peak areas for $2M_1$ and $1M$, respectively were used, as proposed by Grathoff and Moore (1996). The presence of $1M_d$ illite was detected by the presence of the illite hump around the illite 003 diffraction peak (Grathoff and Moore, 1996). WINFIT decomposition by profile fitting was used for determination of areas of the specific peaks of polytypes. Polytype quantification errors are estimated at about ±5%.

The illite crystallinity (IC), or Kübler index, is defined as the width of the first order illite basal reflection (10Å peak) at half height and expressed in Δ2θ values. The Kübler index decreases with increasing illite crystallinity (a measure of the ordering/ thickness of illite crystallites), with temperature being the most important controlling factor (Ji and Browne, 2000 and references therein). IC is a well-accepted mineralogical indicator of the grade of diagenesis, hydrothermal and low-temperature regional metamorphism (Merriman and Peacor 1999; Merriman and Frey, 1999; Ji and Browne, 2000). However, chlorite crystallinity (ChC) or Arkai Index is becoming an additional or alternative technique (particularly in mafic rocks) to evaluate paleotemperature conditions (Árkai, 1991; Warr and Cox, 2016). The ChC is determined through measurement of chlorite 002 peak width (Arkai, 1991). The IC and ChC results of this study were calibrated against the Crystallinity Index Standard (CIS) scale using the procedure and interlaboratory standards of Warr and Mählmann (2015).

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3.2. Petrographic analysis

The thin sections were first examined under plane-polarized light and cross-polarized light conditions using a Nikon Eclipse LV100N POL and a Zeiss Axio Imager.A2m polarizing microscope polarizing microscope. Further examination of the thin sections was undertaken using a Philips XL 40 scanning electron microscope (SEM) equipped with a X-ray energy-dispersive spectrometry (EDS) system for chemical spot analyses. The sections were analysed using 30 kV accelerating voltage and a working distance of 12 mm. Images were collected in back-scattered electron mode. Additionally, clay separates were carbon coated and examined using an EDS equipped Zeiss Ultra Plus SEM qualitative phase identification. The samples were analysed under high vacuum with a 15 kV accelerating voltage and a working distance of 6 mm. Images of clay separate were collected in secondary electron acquisition mode.

3.3. Rb–Sr illite dating

For the Rb–Sr dating (conducted at the Radiogenic Isotope Facility laboratory RIF, the University of Queensland (UQ)), illitic clay separates were leached for 15 min at room temperature in 1 N distilled HCl (Clauer et al., 1993). Leachate and residue were separated by centrifuging. The residue was rinsed repeatedly with milli-Q water, dried and reweighed. Acid leached residues and untreated samples were measured directly by Thermo X-series 1 quadrupole ICP–MS with precision better than 0.5% (1σ). The Sr-enriched fraction was separated using cation exchange resins. Sr isotopic ratios were measured on a VG Sector-54 thermal ionisation mass spectrometer (TIMS. Sr was loaded in TaF_5 and 0.1 N H_3PO_4 on a tantalum or tungsten single filament. Sr isotopic ratios were corrected for mass discrimination using $^{86}Sr/^{88}Sr = 0.1194$. Long-term (6 years) reproducibility of statically measured NBS SRM 987 (2σ; n = 442) is $^{87}Sr/^{86}Sr = 0.710249 \pm 0.000028$. More recent dynamically measured

SRM 987 had $^{86}Sr/^{88}Sr$ ratios of 0.710222 ± 0.000020 (2σ; *n* = 140). Rb–Sr isochron ages were calculated using the ISOPLOT program (Ludwig, 2012).

3.4. K–Ar illite dating

The K–Ar dating was performed at the CSIRO Argon facility in Perth, Australia according to standard methods given in detail by Dalrymple and Lanphere (1969). Potassium content was determined by atomic absorption. The error of K determination of standards is better than 2 %. The K blank was measured at 0.50ppm. Argon was extracted from the separated mineral fraction by fusing the sample within a vacuum line serviced by an on-line ^{38}Ar spike pipette. The isotopic composition of the spiked Ar was measured with a high sensitivity, on-line, VG3600 mass spectrometer. The ^{38}Ar was calibrated against standard biotite GA1550 (McDougall and Roksandic, 1974). Blanks for the extraction line and mass spectrometer were systematically determined and the mass discrimination factor was determined periodically by airshots (small amounts of air for $^{40}Ar/^{36}Ar$ ratio measurement). During the course of the study, 16 international standards (8 HD-B1 and 8 LP-6) and 16 airshots (small amounts of air for $^{40}Ar/^{36}Ar$ ratio measurement) were analyzed. The results are summarized in Table 2. The error for the $^{40}Ar/^{36}Ar$ value of the airshot yielded 296.08 ± 0.23. The general error for argon analyses is below 1% (2σ) based on the long-term precision of Ar measurements of the international standards. The K-Ar age was calculated using ^{40}K abundance and decay constants recommended by Steiger and Jäger, (1977). The age uncertainties take into account the errors during sample weighing, $^{38}Ar/^{36}Ar$ and $^{40}Ar/^{38}Ar$ measurements and K analysis.

3.5. ^{40}Ar–^{39}Ar illite dating

Four fault gouge illites were dated by the ^{40}Ar–^{39}Ar method at the University of Michigan. Illitic clay samples were re-suspended in 1 ml of deionized water, spun-down at

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10,000 rpm in a microcentrifuge and carved into a ~1 mm pellet following decanting. To avoid loss of ^{39}Ar due to recoil, clay pellets were placed in 1 mm ID fused silica vials prior to being sent for neutron irradiation for 90 MWh in medium flux locations of the McMaster Nuclear Reactor (hole 8C for irradiation 1, 8A for irradiation 2). Following irradiation, samples were attached to a laser fusion system, the vials were broken under a 1 x 10–8 Torr vacuum, and the samples step-heated in situ using a defocused beam from a 5 W Coherent Innova continuous Ar-ion laser operated in multi-line mode. Argon isotopes were then analyzed using a VG1200S mass spectrometer equipped with a Daly detector operated in analogue mode using methods by Hall (2014). Ages in this study are calculated relative to an age of 520.4±1.7 Ma for standard hornblende MMhb-1 (Samson and Alexander, 1987). The total gas age obtained from the vacuum encapsulated sample is equivalent to a conventional K–Ar age and quoted at 1σ.

3.6. Illite trace element analysis

For trace element analysis conducted in the Radiogenic Isotope Laboratory at the University of Queensland (RIF, UQ), clay samples were dissolved with a mixture of HF and nitric acids on a hotplate, then evaporated to dryness, refluxed twice with nitric acid and dissolved in 2N nitric acid. Aliquots of the solutions were spiked with internal standards, diluted and analysed on a Thermo X-series 1 quadrupole inductively coupled plasma mass spectrometer (ICP-MS). Sample preparation and analytical procedures used were similar to those of Eggins et al. (1997), except that Tm was not used as an internal standard and duplicate low-pressure digestions of US Geological Survey W-2 diabase standard and a known concentration profile (pre-analysed by laboratory) were used for calibration (Li et al., 2005) were used as the calibration standard. . The ^{156}CeO/^{140}Ce ratio for the run was 0.016. Long-term precision (RSD) was based on duplicate analyses of the duplicate digestions of AGV1, whilst precision for the run was based on five duplicate analyses of W-2 which were better than

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3% for most elements, except for Li, Zn, Mo, Cd, and Cs, which ranged between 5% (Li, Cd and Cs) and 15% (Zn).

4. RESULTS

4.1 Sample description and micro structures

4.1.1 Core descriptions

The undifferentiated Millungera sequence intersected in Julia Creek 1 and Dobbyn 2 comprises medium to coarse-grained, pink to dark red quartzose sandstone with minor interbeds of micaceous clay siltstone and claystone. These sandstone intervals are fractured and faulted throughout the sequence and show evidence of pervasive hydrothermal alteration, particularly near the cracks (Fig. 2a, see also Faulkner et al., 2012; Fitzell et al., 2012). Alteration products are very fine-grained clay-rich material that contain angular clasts from the main rock (Fig. 2b). Clay-rich layers show mostly different colour (grey-beige-red) relative to the sandstone wallrock (Fig. 2b-e). Numerous open cracks coated with green clay are observed throughout cores (Fig. 2c). The clay rich material occurs along the fault planes and in cracks as single vein or complex network of partially consolidated material (Fig. c-d). They also exist as relatively thick layers (up to 30 cm) within the sandstone cores (Fig. 2e), with a sharp transition the host rock (Fig. 2a-f) and contain commonly slickenside surfaces at the contact with the host rock (Fig. 2f-h).

4.1.2 Petrographic and micro-structural analysis

Thin section photomicrographs and SEM images of representative samples are shown in Fig. 3a-f. Microscopic observations show that the undeformed host rock sandstones consist of mainly quartz, some muscovite and minor K-feldspar. Petrography, in combination with XRD analysis show that kaolinite, illite, and chlorite are present as a pore and fracture-filling

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cement in the sandstones (Fig. 3a-b), while detrital mica occurs in large elongate grains with alteration in illite along its edges (Fig. 3b). Chlorite does not show any coarse detrital grains and only occurs authigenetically in very fine-grains dispersed and mixed with illites as pore-filling mineral phases (Fig. 3a).

Faulted specimen from Julia Creek 1 show a characteristic S-C foliation (e.g. Berthe et al., 1979) under the optical microscope (Fig. 4a, b and c) with an anastomosing network of phyllosilicate defining the C shear oriented parallel to the shear direction. The S- shears include planes of insoluble minerals oriented oblique to the sense of shear (Fig. 4c) and quartz fragments embedded in a fine-grained illite-rich matrix as shown by electron microscopy imaging (Fig. 4d and e). These quartz grains have angular shape with intensely serrated grain boundaries and are slightly elongated with their long axis parallel to the orientation of the S- surface. Booklets of kaolinite partially replaced by illite are also visible in the deformed specimens (e.g. Fig. 4d and e).

Faulted hand specimens from Dobbyn 2 exhibit planar to slightly arcuate fault surfaces with a high gloss and display evident slickenside surfaces decorated by short-wavelength (200-500 μm) striations (Fig. 5a). Sense of shearing and offset on the faults are difficult to assess due to lack of markers visible in the cores. Domains of foliated and brecciated cataclasite can be distinguished on hand specimen (Fig. 5b) and in thin section (Fig. 5c). Both domains are characterised by hematite rich injection veins emanating from the slip surfaces and oriented at approximately right angle to them with sharp contacts with the surrounding material (Fig. 5 d, e). The domains and are bounded by sharp contacts defined by slickenside surfaces constituted by thin layers (50-100 μm thick) of iso-oriented phyllosilicates (Fig. 5d, f). The foliated cataclasite domains are characterised by a set of conjugate shears referred to as S-C-C' structures visible at the micro-scale using scanning electron microscopy (Fig. 5e).

Oblique to the shear direction S surfaces are defined by the preferred alignment of elongated clasts of phyllosilicates and are oriented approximately perpendicular to the maximum flattening of the strain ellipsoid.

C and C' represent discrete shear surfaces, the former is parallel to the macroscopic slip surface and the latter deflects the S foliation by disrupting the grains into an ultrafine grained oriented at a small angle (~ 20°) to the macroscopic shear surface but with the opposite sense of obliquity relative to the S-surfaces (Fig. 5e). The cataclasite domain show the original rock fabric of detrital quartz grains and pore-filling diagenetic kaolinite disrupted by a pervasive network of hematite filled intragranular micro-fractures (Fig. 5g).

4.1.3. XRD and SEM clay mineral analysis

Illite is the most abundant clay mineral in the majority of samples, with kaolinite and chlorite being present in many samples. The latter minerals are more abundant than illite in sample Dob-449.3 (Table 1; Supplementary Fig. S1). XRD analysis shows that 001 peak position of the illite does not change after ethylene glycol treatment, which indicates that the amount of smectite-like clays is insignificant (Srodon and Eberl, 1984). IC measurements for <2 μm size fractions normalised to the standards of Warr and Rice (1994) range from 0.17 to 1.00 $\Delta 2\theta$ and from 0.46 to 1.01 $\Delta 2\theta$ for samples from Dobbyn 2 and Julia Creek 1, respectively (Table 1). We also measured IC values of >2 μm size fractions of some fault gouge samples. Such non-clay fractions contain mostly parallel- oriented mica-type inherited/detrital minerals representing the pre-fault protolith to compare their metamorphic grade with that of the fault gouges. Coarser 2 μm size fractions of samples JC-408, Dob-389.6, Dob-441, Dob-449.3, and Dob-476.6 IC give values of 0.42 $\Delta 2\theta$, 0.36 $\Delta 2\theta$, 0.14 $\Delta 2\theta$, and 0.26 $\Delta 2\theta$, respectively. Smaller 2 μm size fractions of the host rock IC values range between 0.20 and 0.27 $\Delta 2\theta$, and 0.46 and 0.51 $\Delta 2\theta$ for Dobbyn 2 and Julia Creek 1 samples,

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respectively (Table 1). The normalised chlorite crystallinity values (ChC) of <2 µm for samples free of kaolinite range from 0.35 to 0.42 *Δ2θ* (Table 1).

Non-oriented random powder XRD analysis of <2 µm, 2-1 µm, <1 µm, and <0.5 µm fractions for samples from borehole Julia Creek 1 (JC) confirm the mixture of 2M, 1M, and1M_d polytypes of illite, while samples from borehole Dobbyn 2 consist of largely 2M illite with some 1M_d illite up to 20% for some samples (Table 1, Supplementary Fi. S1). SEM analysis show 2M illites forming large euhedral crystal plates with sharp edges that occur together with smaller 1M/1M_d illite plates (Fig. 3c-f). A number of previous studies (e.g., Clauer and Liewig, 2013) showed that detrital illitic clay particles rarely have straight edges like those presented in this study, but rather occur in particles with diffuse-blurred and irregular edges (Fig. 3c, the white material on the right-hand site). Samples Dob-441 and Dob-476.6 have generally larger crystal size (Fig. 3c-d) than samples JC-408 and JC-360.7 (Fig. 3c-d). The abundancy of 2M illite represented by these larger crystal plates in samples Dob-441 and Dob-476.6 is confirmed by XRD random powder polytype analysis (Table 1; Supplementary Fig. S1). Dob samples however, are poorly sorted in terms of crystal size distribution with the presence of a number of much smaller crystals (Fig. 3c-d). Such small crystals are mostly rounded (see the arrow in Fig. 3d) and probably represent the 1M_d illite recorded in random powder XRD pattern of sample Dob-441. It is apparent that larger crystal plates were formed by coalescence of smaller crystals through Ostwald ripening (Eberl et al., 1990) (e.g., dashed arrows in Fig. 3d).

4.2. Illite geochronology

4.2.1. ^{40}Ar–^{39}Ar dating

Four fault gouge illite samples of <2µm fraction were analysed for ^{40}Ar–^{39}Ar geochronology (Table 1 and Fig. 6). Based on their illite crystallinity values, these samples

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represent high diagenetic to upper anchizonal metamorphic grade with 2M illite varying between 62% and 100%. Samples had 0.050%–0.12% low temperature ^{39}Ar recoil loss, which is characteristic of well crystallised illite grains (Hall et al., 1997). Age data (1σ) is obtained as total gas ages (Table 1 and Fig. 6) (cf., Dong et al., 1995). Samples JC-360.7, JC-408, Dob-441, and Dob-446.6 yield total gas of 1038.1 ± 2.9 Ma and 1099.6 ± 3.0 Ma, 1040.0 ± 2.3 Ma and 1106.8 ± 2.5 Ma (Table 1; Fig. 6). The analyses do not show well developed plateaux, which can be explained by recoil and varying ages of individual crystals (cf., Clauer et al., 2012).

4.2.2. K–Ar dating

K–Ar ages of fault gouge and sandstone illites of different size fractions from >2 μm to <0.1 μm from boreholes Julia Creek 1 and Dobbyn 2 are presented in Table 3 and Fig. 7. A histogram of all K–Ar results obtained from gouge zones are shown in Fig. 7a. K–Ar size fraction ages for fault gouge and host rock matrix illite and their interpretation in relation to the tectonic history are shown in Fig. 7b and Fig. 7c, respectively. Due to sample nature it was not possible to extract sufficient material for <0.1 μm or <0.5 μm fractions from some fault rock samples, especially from those samples from Dobbyn 2 with illites with low IC values (0.42 *Δ2θ* or lower for <2 μm).

Size fractions from <2 μm to <0.1 μm of fault gouge samples JC-343, JC-360.7, and JC-440.5 from Julia Creek 1 yield consistent ages (Table 1) with a mean (average) of 1036.2 ± 9.7 Ma, 1025 ± 17.7 Ma, 1028.9 ± 17.3 Ma (1 s.d.), respectively. The mean age of 1025 ± 17.7 Ma for sample JC-360.7 is identical with the Ar–Ar total gas age of 1038.1±2.9 Ma of <2 μm of the same sample. Various size fractions from 2-1 μm to <0.5 μm of another fault gouge sample from Julia Creek 1 (JC-408) give also consistent but older ages with a mean of 1114.2 ± 6.9 Ma. However, ^{40}Ar–^{39}Ar total gas age of sample JC-408 is 1040.0 ± 2.3 Ma,

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which is identical to the above mentioned ages of JC-343, JC-360.7, and JC-440.5 samples (Table 1; Fig. 7a). The >2 μm fraction of sample JC-408 yields, by contrast, a distinctively different and older K–Ar of 1243.2 ± 29.1 Ma (Table 1; Fig. 7a).

K–Ar ages of different fault gouge illites from Dobbyn 2 are more variable (Fig. 5a). 2-1 μm, <2, 1-0.5 μm, and 0.5-0.1 μm fractions of sample Dob-389.6 yield consistent ages (Table 1; Fig. 7a) with a mean of 1061 ± 19.5 Ma, whereas <0.5 μm and <0.1 μm give younger ages of 981.8 ± 22.6 Ma and 905.4 ± 20.9 Ma, respectively.

Smaller 2 and <1 μm fractions of fault sample Dob-441 give inconsistent but close K–Ar ages of 1148.7 ± 26.9 Ma and 1086.5 ± 25.1 Ma, respectively. The ^{40}Ar–^{39}Ar total gas age of 1068.1 ± 1.8 Ma for this sample is consistent with the K–Ar age of 1086.5 ± 25.1 Ma of the <1 μm fraction. A significantly older K–Ar age of 1312.3 ± 30.7 Ma is obtained for the >2 μm fraction of sample Dob-441 (Table 1; Fig. 7a).

Samples Dob-449.1 and Dob-449.3 were taken from a clay-rich fault rock zone, with the former and latter representing beige-light grey and hematite-rich red varieties, respectively. The 2-0.5 μm, <2 μm, 1-0.5 μm, and <0.5 μm fractions of sample Dob-449.1 yield younger but concordant ages with a mean of 922.4 ± 19.9 Ma. However, illite fractions of sample Dob-449.3 (just 20 cm below) yield scattering K–Ar ages regardless of the grain size. K–Ar ages of size fractions 2-1 μm, 1-0.5 μm, 0.5-0.2 μm, and <0.5 μm for sample Dob-449.3 are 1047.7 ± 24.2 Ma, 1117.2 ± 25.8 Ma, 950.9 ± 22.0 Ma, and 1004.4 ± 23.2 Ma, respectively. Coarser 2 μm fraction being rich in detrital mica gives a much older age of 1259.0 ± 29.1 (Table 1; Fig. 7a).

K–Ar age of 2-1 μm fraction and K–Ar and ^{40}Ar–^{39}Ar total gas ages of <2 μm fraction of the deepest fault rock sample from Dobbyn 2 (Dob-476.6) yield identical ages within

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analytical errors of 975.7 ± 22.2 Ma, 983.7 ± 23.0 Ma, and 994.6 ± 2.2 Ma, respectively, with a mean of 984.7 ± 9.5 Ma. A much older age of 1170.4 ± 27.4 Ma is obtained for >2 μm fraction of this sample (Table 1; Fig. 7a).

K–Ar and ^{40}Ar–^{39}Ar results of all size fractions (except >2 μm) from fault gouges listed in Table 1 are presented as a histogram and probability density distribution plot (Fig. 7b). Isotopic dates define distinct age clusters at ~1070 Ma, ~1040 Ma, and ~995Ma. There are also less pronounced, but noticeable age clusters at ~1115 Ma and ~905 Ma (Fig. 7b).

K–Ar ages of different size fractions of illitic clay minerals that occur as matrix in undeformed, adjacent sandstones (see Figs. 2 and 3) are also presented in Table 1 and Fig. 7c. Three different size fractions of three different samples, Dob-449.4, JC-500, and JC-360.6, yield same ages within error, averaging at 1047 ± 21 Ma, 1079 ± 13 Ma, 1062 ± 7 Ma (1 s.d), respectively (Table 1 and Fig. 7c). The <0.2 fine fractions of JC-360.6 and JC-500 yield within error identical younger agers of (928.3 ± 47.6 and 878.3 ± 45.1, respectively), which might indicate cessation of illite formation or partial reset due to final faulting with the Early Neoproterozoic deformation events and associated fluid flow.

4.2.3. Rb–Sr isochron dating

Rb–Sr data for the untreated, acid-leached residues, and leachates of different <2μm clay fractions for the fault gouge illites collected from different stratigraphic levels in Julia Creek 1 and Dobbyn 2 are presented in Table 4 and on Fig. 8. The data show three parallel well-defined linear relationships indicating similar isochron ages, but with different initial ^{87}Sr/^{86}Sr values (Fig. 8a). Some samples plot off these lines, possibly because they have different initial ^{87}Sr/^{86}Sr values, and these samples are not considered for isochron age calculation. Samples from Dobbyn 2 plot on the two upper isochron lines with higher ^{87}Sr/^{86}Sr initial values (Fig. 8a). Residue of samples JC-360.7B<0.5 μm plot also on one of these lines (the middle line on Fig. Fig. 8a). All other Julia Creek 1 define a separate Rb–Sr isochron line with lower $^{87}Sr/^{86}Sr$ initial values (the lower line on Fig. 8a).

Leachates are accessory acid-soluble non-silicate phases (mostly carbonate minerals and amorphous grain coatings of FeO(OH) (Clauer et al., 1993). However, Rb–Sr isotopic systematics of the acid-soluble leachate is not in equilibrium with that of the illites, since the leachates plot off the Rb–Sr lines. Lower $^{87}Sr/^{86}Sr$ values of the leachates (mostly<0.72) in comparison to highly radiogenic (elevated) initial $^{87}Sr/^{86}Sr$ of of illites indicate interaction of rocks with some late stage fluids from which acid-soluble non-silicate phases were formed.

The data of untreated and residues of <2 μm fractions from Julia Creek 1 samples define a linear relationship from which the slope yields a Rb-Sr errorchron age of 1041 ± 46 Ma (initial $^{87}Sr/^{86}Sr$ = 0.7194±0.0011, MSWD=27) (Fig. 8b). However, as apparent from the $^{87}Rb/^{86}Sr$ vs. $^{87}Sr/^{86}Sr$ plot on the Fig. 8a-b, untreated aliquots of samples JC-387.8 and JC-440.5 plot slightly at the lower part of the line leading to a large analytical error and MSWD value. This is probably caused by the effect of the leachable components that were not in isotopic equilibrium with the clays, as discussed above. When these two untreated samples are omitted, the data scatter is reduced significantly with a well-defined regression line (MSWD=2.3) and corresponding isochron age of 1023 ± 12 Ma (initial $^{87}Sr/^{86}Sr$ = 0.72009±0.00025) (Figure 8c). Residue and untreated aliquots of <2 μm fractions from Dob-449, Dob-441A, and Dob-389.6, and residue of JC-360.7B <0.5 μm yield an analytically indistinguishable age of 1033 ± 25 Ma (initial $^{87}Sr/^{86}Sr$ = 0.72326±0.00094; MSWD=2.5) (Fig. 8d). A somewhat younger Rb–Sr age of 1000 ± 12 (initial $^{87}Sr/^{86}Sr$ =0.72841±0.00030, MSWD = 0.065) was obtained for Dob-476.6 <2 μm (untreated and residue), Dob-389.6 2-1 μm and 0.5-0.1 μm (residue) (Fig. 8e).

4.3. Trace elements

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Rare earth element (REE) data and Th, U, and Sc contents of illites (<2 μm clay-size fractions) from the fault gouge samples are given in Table 4. Chondrite-normalised REE patterns of illites from the fault gouges are shown in Fig. 9a. In addition, the REE pattern of Post-Archean Average Shale (PAAS, Taylor and McLennan, 1985) is included in the REE diagram. The fault gouge illites are substantially enriched in light REE (LREE) relative to PAAS with La contents as high as10xPAAS. The illites are however, somewhat depleted in heavy REE (HREE) relative to LREE (Fig. 9a). The chondrite-normalised $(La/Lu)_c$ ratios of the illites are significantly higher (up to 76) than the $(La/Lu)_c$ ratio of PAAS (10) (Table 4). Fault gouge illites are also enriched in Th and U (up to 10 times) in comparison to PAAS (Table 4).

5. DISCUSSION

5.1. Faulting, fluid-rock interactions and clay generation

Brittle deformation and faulting is evident from cores in the sampled intervals in Julia Creek 1and Dobbyn 2, probably associated with the large scale faults inferred from interpretations of seismic survey (Fig. 1). Under upper crustal conditions, fault zones accommodate intense shear strain often localised in bands of cataclastic deformation formed by friction-dominated faulting within the seismogenic regime (Sibson, 1977; Schmid and Handy, 1991). While cataclastic fault rocks are generally considered to display random fabric, foliated fault rocks such as fault gouge and foliated cataclasites have also been reported in different lithologies ranging from crystalline rocks to siliciclastic and carbonate dominated sediments (Chester et al., 1985; Rutter et al., 1986; Lin, 1999; Ujiie et al., 2007; Laurich et al., 2004; Delle Piane et al., 2017; Nicchio et al., 2018). Frictional sliding and abrasion are common processes during repeated fault movement and result in strong grain size reduction of the fault rocks with respect to the constituting minerals in the undeformed portion of the host rocks. The abundant

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presence of micro and nano-sized particles in cataclasites and gouges may result from combined effects of cataclasis and pressure solution-precipitation during deformation in the presence of fluids (e.g. Vrolijk and van der Pluijm, 1999; Solum et al., 2005). The corroded grain boundaries of quartz grains in faulted samples from Julia Creek 1 (Fig. 4) and the presence of the injection veins and hydrothermal hematite in the cataclasites from Dobbyn 2 (Fig. 5) indicates that deformation occurred in a fluid-rich environment that promoted detrital muscovite dissolution and new growth of illite. The small injections veins that are observed to cut through the foliated cataclasites and the detrital quartz grains (Fig. 5) may represent the effect of hydraulic fracturing due to a fast increment of fluid pressure in the fault zone during a seismic slip (e.g. Sibson, 1989; Cowan et al., 2003; Ujiie et al., 2007; Rowe et al., 2012). At the core scale, some samples show no shearing-related fabrics in the sandstone cores (fresh and hard) adjacent to clay-filled cracks (Fig. 2a-d). This may be a result of the precipitation of clay-rich material and injection of granular material from seismically-mobilised circulating fluids.

K–Ar results show that illites from fault gouges and matrix illites in undeformed adjacent sandstones precipitated contemporaneously (Fig. 7). In some tectonically active regions, mineral assemblages from the fault rocks and their parent rocks are significantly different, whereby parent rocks do not contain any alteration minerals with new mineral growth being restricted to the fault rocks. This indicates that the heat and fluid flows associated with mineral authigenesis were not controlled by regional tectonic events in these regions, but rather confined to the areas within the fault zone (e.g., Uysal et al. 2006; Isik et al., 2014; Babaahmadi et al., 2019). However, the relation between large-scale fluid flow and seismic events has long been reported (e.g., Bruhn et al., 1994; Eichhubl et al., 2010; Faulkner et al., 2010; Lupi et al., 2010 and references therein). Brittle faulting in the upper crust involves episodic changes in the stress level that can expel large volumes of fluids, leading to the

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generation of hydrothermal/geothermal systems. Faults and veins and their immediate surrounds represent zones of fluid passage and transfer of mass through those fluids (e.g., Sibson, 1987). Mineral alteration in slip zone gouge extends outward from the fault zone into the undeformed wall rock (e.g., Parry et al., 1991; Craw et al., 2009). The wall rock alteration is attributed to the diffusion and advection of fluids, and hence chemical mass and heat transfer associated with deformation. For example, metasomatic alteration zones develop around fluid pathways by advection with mineral dissolution and precipitation increasing towards the conduit and dictated by infiltrating fluids (Ferry and Dipple, 1991). Metasomatic mineral alteration is common in sedimentary basins contemporaneous with regional extensional tectonics. Alteration is driven by reactivity of sandstone host rocks with illitic clay minerals, K-feldspar (adularia), hematite, calcite, and quartz being some common minerals precipitating from tectonically mobilised K-bearing basin brine (Fedo et al., 1995; Michalski et al., 2007; Eichhubl et al., 2010). Similarly, in this study, sandstone immediately above or below the fracture and shear (fault gouge) zones represents zones of alteration of detrital minerals to illite by tectonically moved fluids.

5.2. Geochronology: comparison between Rb–Sr, K–Ar and ^{40}Ar–^{39}Ar ages

Fault gouges from the Millungera Basin in Australia (Fig. 1) contain a mixture of coevally formed 1M/1M$_d$ and 2M illite from which the crystallisation age was determined by a combined application of Rb–Sr, ^{40}Ar–^{39}Ar, and K–Ar techniques. It is demonstrated that applying both the Rb–Sr and K–Ar (^{40}Ar–^{39}Ar) techniques for dating the same fault gouge minerals provides more robust and complementary age constraints on faulting episodes and minimises the inherent disadvantages for each isotopic system. A common drawback of Ar geochronology when dating white mica is that Ar apparent ages are either significantly older or younger than the Rb–Sr isochron ages of the same samples (Kelley, 2002; Di Vincenzo et

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al., 2006). Rb–Sr isotopic systematics may remain unaffected because Rb–Sr resetting requires higher closure temperatures and sufficient fluids in the system to facilitate recrystallisation (e.g., Di Vincenzo et al., 2006). However, a potential pitfall of Rb–Sr dating technique could result from heterogeneous initial $^{87}Sr/^{86}Sr$ ratios on a mineral-scale (cf., Davidson et al., 2005). In this study, as discussed in section in section 4.2.3, we minimised the effect of mineral-scale initial isotopic heterogeneity by analysing different aliquots (untreated, leachates, and residues) and different sub-size fractions of one clay sample. Indeed, different sub-size fractions and aliquots of some samples plot on different isochron lines indicating different $^{87}Sr/^{86}Sr$ ratios (Fig. 8, see the discussion below).

Illites from the Millungera Basin fault gouges display well-developed linear data arrays on Rb–Sr isochron diagrams with statistically valid late Mesoproterozoic ages (Fig. 8). Such linear relations for fault gouge clays representing acid-leached residue and untreated aliquots and consisting of both 2M and 1M illite can be interpreted as unequivocal indication of a complete isotopic equilibration of the entire illite population during episodic fluid flow episodes controlled by deformation events (cf. Mutlu et al., 2010; Isik et al., 2014; Middleton et al., 2014; Rosenbaum et al., 2015; Babaahmadi et al., 2019). The Rb–Sr isochron ages (1041 ± 46 Ma and 1023 ± 12 Ma) for <2 μm fractions of samples JC-360.7, JC-440.5 (Fig. 8b-c) are consistent with K–Ar and for $^{40}Ar–^{39}Ar$ (sample JC-360.7) ages of the same fraction and all other different size fractions of the same samples (Table 1 and Fig. 7). Similarly, the Rb–Sr isochron age of 1033 ± 25 Ma for <2 μm fraction of sample Dob-389.6 (Fig. 8d) is concordant with the same size fraction, and 1-0.5 μm and 0.5-0.1 μm fractions of this sample (Table 1 and Fig. 7). Smaller 2 μm fractions of samples Dob-441A and Dob-449.1 in addition plot on the same Rb–Sr isochron in Fig. 8d, with consistent $^{40}Ar–^{39}Ar$ age (1068.1 ± 1.8 Ma) and K–Ar ages of <1 μm and <0.5 μm fractions of Dob-441. However, <2 μm of Dob-441 yield an older K–Ar age (1148.7 ± 26.9 Ma) (Table 1 and Fig. 7). Unlike the Rb–Sr isochron age (Fig. 8d), K–Ar ages of all different size fractions for sample Dob-449.1 are lower (925 to 913 Ma, Table 1 and Fig. 7). The Rb–Sr isochron age of 1000 ± 12 Ma for <2 µm of sample Dob-476.6 (along with the acid-leached residues of Dob-441B and Dob-389.6 for 2-0.5 µm to 0.5-0.1 µm fractions) is consistent with K–Ar ages of the same sample for various size fractions (922.2 ± 21.2 to 983.7 ± 23.0 Ma) within analytical errors (Table 1 and Fig. 7).

In summary, most samples with their various size fractions yield consistent Rb–Sr isochron and individual K–Ar (^{40}Ar–^{39}Ar) ages. This finding is similar to those reported by some recent studies that presented robust and comprehensive Rb–Sr, ^{40}Ar–^{39}Ar, and K–Ar age data for high diagenetic and anchizonal fault gauge illites (e.g., Middleton et al., 2014; Rosenbaum et al., 2015; Babaahmadi et al, 2019). Discrepancy in Rb–Sr and K–Ar ages for few samples, e.g., Dob-441 <2 µm (K–Ar age is older) and Dob-449.1 (K–Ar ages are younger), may result from sample heterogeneity (hand-specimen and micro scale, see Figs. 4 and 5) presented by more than one illite generation due to multiple faulting episodes. Mixing of different generations is possible in small scale as a result of variable degrees of isotopic resetting of a single illite generation during subsequent faulting events. Pervasive overprinting and re-crystallization can be hindered even in micro scale, which may result from a lack of permeability and/or limited availability of fluids (e.g., Bröcker et al., 2013).

5.3. Implication of Sr isotope and trace element compositions for the evolution of fault-related fluids

The trace element composition of authigenic clay minerals reflects the mineral/fluid partition coefficients for different elements, as well as the composition of fluids from which the clays precipitated. Trace element contents and concentrations, as well as some element ratios, can be used to trace the origin of basinal and hydrothermal fluids (e.g., Uysal and Golding, 2003; Uysal et al., 2005; Uysal et al., 2011). The fault gouge illites analysed in this

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study are highly enriched in LREE and other incompatible elements such as Th and U relative to PAAS (Fig. 9a and Table 4). This geochemical characteristic indicates precipitation of the illites from fluids that must have interacted with rocks of the upper crust enriched in incompatible and heat producing elements. This is also consistent with initial $^{87}Sr/^{86}Sr$ values of fault gouge illites that reflect Sr isotope composition of fluids from which the illites precipitated. The radiogenic initial $^{87}Sr/^{86}Sr$ ratios of about 0.72 indicate the involvement of fluids that equilibrated with old Rb-rich crustal rocks. The trace element and Sr isotope data are in agreement with seismic and potential field data by Korsch et al. (2011), which is interpreted as indicating the occurrence of granites with a thickness up to 5.5 km below the Millungera Basin. The inferred granites may be a part of the granite (Williams Supersuite) exposed just to the west in the Mt Isa Inlier, which is enriched in Th, U, and K (Korsch et al., 2011 and references therein).

Different $^{87}Sr/^{86}Sr$ initial values of illites of the parallel isochron lines corresponding to the same Rb–Sr age in Fig. 8 and scatter of Rb–Sr data points for some samples indicate separate circulation pathways for seismically mobilised fluids that that might have restricted to unconnected fault planes and fracture systems in different areas of the Millingera Basin.

5.4. Significance for regional tectonics

Northeast Australia lies on a cratonic margin that has had a complex crustal history involving the successive development of several Proterozoic to Paleozoic orogenic systems (Fig. 10). Age data from the faults defining the margins of the Millungera Basin is thus important in revealing concealed major Proterozoic tectonic zones in Australia, which contain energy and mineral resources (Korsch et al., 2011).

The fault gouge ages clustering at (~1115 ± 26 Ma, ~1070 ± 25 Ma, ~1040 ± 24 Ma, ~1000 ± 23 Ma, and ~905 ± 21 Ma (Fig. 7) may be related to the regional extension and

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associated major thermal event that occurred across Australia at 1120-900 Ma (e.g., Musgrave Orogeny and subsequent Giles Event), due to interactions between Australia and other continents during assembly of the supercontinent Rodinia (De Vries et al., 2008; Li et al., 2008; Evins et al., 2010). This Australia-wide tectono-thermal event that largely developed along former (Mesoproterozoic) collision zones led to emplacement of widespread dyke swarms, sills and associated granite plutons in the central and north Australia craton largely in a time frame between ~1040 Ma and ~1090 Ma (Schmidt et al., 2006; Evins et al., 2010; Aitken et al., 2013). The Musgrave Orogeny involving widespread emplacement of granite and mafic–ultramafic bodies were recorded in central Australia at 1220 Ma and 1120 Ma (Evins et al., 2010; Kirkland et al., 2013). Major swarms of dolerite intrusions in the North Australian Craton dated at 1116±12 Ma (Lakeview Dolerite, Tanaka and Idnurm, 1994) and associated hydrothermal events were recorded in the Mt. Isa Province (adjacent to the study area, Fig. 1) (Uysal et al., 2004). The illite ages clustering around ~1100 Ma coincide with the latest stage of the Musgrave Orogeny Fig. 7b, Fig. 10).

Another cycle of mafic intrusions in central Australia occurred during the extensional Giles Event between ~1078 Ma and 1068 Ma (Evins et al., 2010; Aitken et al., 2013 and references therein), which was followed by granite magmatism and accompanied felsic volcanism between ~1050 Ma and ~1040 Ma (Evins et al., 2010 and references therein). The latest phase of the Giles Event is represented by the felsic Smoke Hill Volcanics yielding a date of 1026±26 Ma. The K–Ar ages clustering around ~1040 Ma of different illite size fractions from the fault gouges and sandstones are consistent with the timing of the later stage of the Giles event.

Orogenic events post-dating the Giles event are represented by mafic dykes and rare pegmatites emplaced at about 1000 Ma (Evins et al., 2010). Further, a Rb–Sr age of 897 ± 9 Ma is reported for dolerite from the Stuart Dyke Swarm in the southern part of the Arunta

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Block, Northern Territory (Black et al., 1980). The Rodinia supercontinent assembled through worldwide orogenic events by 900 Ma. Stresses induced by the ca. 900 Ma event probably caused reactivation of older orogens within Rodinia (Li et al., 2008). The Amadeus Basin in north central Australia was initiated at ~900 Ma in the late Proterozoic by crustal extension, probably in associations with mafic intrusions being correlated with the Stuart Dyke Swarm (Korsch and Lindsay, 1989). Fault gouge K–Ar ages of ~900 – 950 Ma for various size fractions from sample Dob-449.1 and finest fraction (<0.1 mm) from Dob-389.6 coincide with the timing of deformation associated with these early Neoproterozoic igneous and deformation events (Fig. 7b).

The dated faults of the Millungera Basin may be associated regionally with a series of fault systems bounding rift basins in the southern Georgina Basin. Those fault zones (e.g., Burke River Structural Belt, Pilgrim Fault Zone, Greene, 2010), which are in close proximity to and run parallel to the dated faults framing the Millungera Basin, occur extensively in the adjacent Mt Isa Inlier (Greene, 2010; Korsch et al., 2011). The Pilgrim Fault Zone was established a Mesoproterozoic structural boundary within the Mt. Isa Inlier (Greene, 2010). The southern Burke River Fault, just to the west of the Millungera Basin (Fig. 1), represents a rift-bounding normal fault, were reactivated and inverted to reverse faults during the mid-Paleozoic Alice Springs Orogeny (~400-350 Ma) (Greene, 2010). Similarly, samples from this study were taken from thrust faults at the margin of the Milungera Basin. However, the ^{40}Ar–^{39}Ar, K–Ar, and Rb–Sr ages of the fault gouge illites have been essentially preserved and no tectonic event after about 905 Ma has reset the isotopic systematics of these fault gouges (Fig. 10). In conclusion, our geochronological age data constrain the timing of fault activity associated with the late Meseproterozoic and early Neoproterozoic emplacement of the intrusions and crustal regional extension in central-north Australia.

6. CONCLUSIONS

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A new integrated study was conducted employing radiometric age dating (Rb–Sr, ^{40}Ar–^{39}Ar, and K–Ar) of illitic clay minerals from fault gouges and Neoproterozoic host sandstones bounding the recently discovered Millungera Basin in north-central Australia. Rb–Sr isochron, ^{40}Ar–^{39}Ar total gas, and K–Ar ages are consistent indicating late Mesoproterozoic and early Proterozoic episodes (~1115 ± 26 Ma, ~1070 ± 25 Ma, ~1040 ± 24 Ma, ~1000 ± 23 Ma, and ~905 ± 21 Ma) of active tectonics in north-central Australia. These faulting episodes correspond to timing of regional extension and associated major thermal event that occurred across Australia at 1120-900 Ma, due to interactions between Australia and other continents during assembly of the supercontinent Rodinia. Sr isotope and trace element data indicate that fault gouge illites precipitated from fluids that interacted with deep granitic basement enriched in heat-producing elements. This study provides insight into the inscrutable time-space distribution of Precambrian tectonic zones in central Australia, which are responsible for the formation of a number of sedimentary basins with significant energy and mineral resources. Investigating core samples with preserved isotopic signature of Proterozoic fault rocks avoids the effect of surface weathering of old geological terranes.

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Figure captions

Figure 1. Simplified map of north-central Australia showing the interpreted subsurface distribution of the Millungera Basin. The surface distribution of Cenozoic and Mesozoic sediments and the locations of the Proterozoic–Ordovician Basins are also shown (modified from Korsch et al., 2011).

Figure 2. Photos of sampled cores showing fault gouges investigated in this study. Foliated and veined clay-rich fault gouges with light grey – green colour distinctive from adjacent brown hematite-rich sandstone host rock (a-e). Fault gouge veins are characterized by an ultrafine-to fine-grained matrix and angular to sub-angular fragments of host sandstone of various sizes, ranging from submicron to centimetres (2b). Ultracataclastic veins are common, which are observed as simple veins, complex lenses, and networks (c-d). Slickenside structure is a commonly seen at the sharp contact between the clay-rich fault gouge layers and the host rock (f-g).

Figure 3. Thin section photomicrographs (a-b) and SEM images (c-f) illustrating fault gouge illites from the Millungera Basis. (a) Illite plates (white) occur in voids within detrital quartz grains and as pore-filling clay between detrital grains together with chlorite. Note green – yellow chlorite shown by red arrow. (b) Alteration of detrital muscovite in illite. Note illite plates at the ends of the mica filling pores. (c) SEM image of sample Dob-441. Note the large detrital mica grain (>2 μm) with diffuse-blurred and irregular edges (the white material on the right-hand site), while authigenic illites occur in smaller crystals (<2 μm) with straight edges. SEM image of upper anchi-and epizone sample Dob-476.6. Note rounded smaller crystals (arrows), which occur partly as a constituent of larger illite plates (dashed arrows). (e-f) SEM images of samples JC-408 and JC-360.7, respectively. Note euhedral (hexagonal) and anhedral crystal plates with sharp and straight edges of these JC samples, which occur in

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smaller crystal size in comparison to the Dob samples. Smaller crystal size is consistent with higher illite crystallinity values of JC samples (see Table 1).

Figure 4. Microstructures of a faulted sample from Julia Creek 1 well (depth 360 m), petrographic thin section is cut parallel to the inferred shear direction. A) Whole thin section image collected using an optical microscope in plane polarised light showing composite S-C foliation (red dashed lines illustrate the orientation of the S plane. B) Optical microscope image in cross polarised light showing phyllosilicate enriched C-planes and the oblique S-foliation. C) Optical microscope image in cross polarised light showing alignment of opaque insoluble minerals along the S-plane (highlighted by the red arrows) indicative of pressure solution. D and E) Scanning electron microscope images of the deformed rock showing corroded boundaries in detrital quartz (Qtz) highlighted by the red arrows, authigenic kaolinite and illite (Kao; Ill) and detrital muscovite (Msv) aligned along the C plane.

Figure 5: Faulted sample from Dobbyn 2 well (depth 441 m). A) Hand specimen showing green-beige slickenside surface with striations due to frictional movement along the surface. B) Polished face of the hand specimen cut parallel to the shear direction as inferred from the striation direction shown in A). C) Whole thin section image collected using an optical microscope in planes polarised light. The position of the thin section with respect to the hand specimen is shown by the red rectangle in B); yellow dashed line bound different microstructural domains of the fault rock defined as i) foliated cataclasite and ii) cataclasite. White boxes indicate the location of the following images. D) Optical microscope image in plane polarised light showing the slickenside surface (bound by the yellow dashed line) of the samples being composed of iso-aligned phyllosilicates. Also shown is a hematite rich injection vein. E) Scanning electron microscope image of the foliated cataclasite portion of the samples and the characteristic S-C-C' texture (see text for details). White dashed line highlights the orientation of the S-planes. F) Optical microscope image in plane polarised

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light showing a slip surface (bound by the yellow dashed line) at the boundary between the foliated and non-foliated cataclasite domains. G) Optical microscope image in plane polarised light showing a network of hematite filled intra-crystalline micro-fractures in the cataclasite domain of the fault rock, mainly composed of quartz (white grains) and pore filling clays.

Figure 6. Ar–Ar dating results and argon release diagrams for illites from the fault gouges of <2 μm size fraction. Note consistent Ar–Ar total gas ages except for sample Dob-476.6.

Figure 7. A) Histogram for K–Ar and Ar–Ar ages and probability distribution of ages for fault gouge illites. Curves show relative probabilities calculated using Isoplot 7 for Excel (Ludwig, 2012). B) K–Ar and Ar–Ar dates (no error bars due to small errors) for different size fractions of fault gouges and C) matrix illites from host rocks and their interpretation in relation to tectonic history.

Figure 8. A) Rb-Sr data of the different size fractions and the untreated, leachate and residue separates of each size fractions from sample. Parallel linear relationships correspond to similar isochron age, but with different initial $^{87}Sr/^{86}Sr$ values. B) Rb-Sr plot for untreated and residues of <2 μm fractions from most Julia Creek 1 samples. C)Well-defined isochron for Julia Creek 1 samples after omitting two untreated aliquots D) Rb–Sr isochron diagrams for an assemblage of Dobbyn 2 samples (including JC-360.7B <0.5 μm) and E) another group of Dobbyn 2 samples with a younger age.

Figure 9. REE patterns of the fault gouge illites. Note that the illites are substantially enriched in light REE (LREE) relative to PAAS.

Figure 10. Summary of the geological history of the Millungera basin region and isotopic age clusters of the illites.

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Table 1. Clay mineralogy and age data of fault gouges and host rocks from Julia Creek 1 (JC) and Dobbyn 2 (Dob).

Sample & size fraction	Clay mineralogy	IC ($\Delta 2\theta$)	ChC ($\Delta 2\theta$)	Very-low-grade metamorphic zone	2M %	1M %	$1M_d$ %	Rb–Sr isochron age	Ar–Ar total gas age	K–Ar age	K_2O (%)	$^{40}Ar_{rad}$ (%)	$^{40}Ar_{rad}$ (10^{-10}mol/g)
Fault gouge													
JC-321 <2mm	illite	0.59		Diagenetic									
JC-326.1 <2µm	illite	0.69		Diagenetic									
JC-343.3 <2µm	illite, kaol., chl.	0.63		Diagenetic	68	14	18			1044.0 ± 24.4	7.10	99.3	174.1
JC-343.3 0.5-0.1µm	illite, kaol., chl.	0.79		Diagenetic						1039.2 ± 24.0	6.86	99.7	167.2
JC-343.3 <0.1µm	illite	0.92		Diagenetic						1025.3 ± 23.7	6.33	99.6	151.6
JC-360.7 <2µm	illite, kaol., chl.	0.63		Diagenetic	66	18	16	1023 ± 12	1038.1 ± 2.9	1014.9 ± 23.7	7.61	99.3	179.8
JC-360.7 2-1µm	illite, kaol., chl.	0.65		Diagenetic	58	22	20			1038.9 ± 24.0	7.49	99.4	182.5
JC-360.7 <1µm	illite, kaol., chl.	0.73		Diagenetic						1041.1 ± 24.1	7.49	99.5	183.0
JC-360.7 <0.5µm	illite	0.75		Diagenetic						1005.5 ± 23.1	7.65	99.6	178.6
JC-387.8 <2µm	illite, kaol., chl.	0.57		Diagenetic									
JC-408 >2µm	illite, kaol., chl.	0.42		Upper anchizone						1243.2 ± 29.1	7.64	99.7	237.1
JC-408 2-1µm	illite, kaol., chl.	0.63		Diagenetic	63	14	23			1118.7 ± 25.9	7.94	99.6	213.4
JC-408 <2µm	illite, kaol., chl.	0.60		Diagenetic	64	17	19		1040.0 ± 2.3	1115.8 ± 26.1	7.88	99.5	211.1
JC-408 <1µm	illite, kaol., chl.	0.70		Diagenetic	64	15	22			1118.2 ± 25.9	7.81	96.6	209.8
JC-408 <0.5µm	illite	0.82		Diagenetic						1104.0 ± 25.4	7.82	99.7	206.5
JC-430.4 <2µm	illite, kaol., chl.	0.51		Lower anchizone									
JC-440.5 <2µm	illite, kaol., chl.	0.44		Lower anchizone	58	24	18	1023 ± 12		1048.9 ± 24.5	6.60	99.4	176.3
JC-440.5 0.5-0.1µm	illite, kaol., chl.	0.84		Diagenetic						1020.3 ± 23 .6	7.41	99.8	168.8
JC-440.5 <0.1µm	illite	1.01		Diagenetic						1017.6 ± 23 .5	7.12	99.8	168.8
JC-473-A <2µm	illite, chl.	0.51	0.35	Lower anchizone									
JC-473-B <2µm	illite, chl.	0.47		Lower anchizone									
JC-483.2 <2µm	illite, chl.	0.60	0.42	Diagenetic									
Dob-389.6 2-1µm	illite, kaol., chl.	0.42		Upper anchizone	95		5			1081.8 ± 25.0	5.97	99.7	153.5
Dob-389.6 <2µm	illite, kaol., chl.	0.43		Upper anchizone	90		10	1033 ± 25		1071.2 ± 25.0	6.53	99.7	165.7
Dob-389.6 1-0.5µm	illite, kaol., chl.	0.51		Lower anchizone	90		10			1037.8 ± 24.0	7.32	99.6	178.1
Dob-389.6 0.5-0.1µm	illite, kaol.	0.62		Diagenetic	80		20			1053.0 ± 24.3	7.20	99.7	178.6
Dob-389.6 <0.5µm	illite, kaol.	0.63		Diagenetic						981.8 ± 22.6	7.63	99.8	172.7
Dob-389.6 <0.1µm	illite, kaol.	1.00		Diagenetic						905.4 ± 20.9	6.62	99.6	135.0
Dob-417 <2µm	illite, kaol., chl.	0.50		Lower anchizone									
Dob-441 >2µm	illite, kaol., chl.	0.36		Upper anchizone						1312.3 ± 30.7	3.72	99.5	124.5
Dob-441 <2µm	illite, kaol., chl.	0.42		Upper anchizone	95		5	1033 ± 25	1068.1 ± 1.8	1148.7 ± 26.9	4.98	99.3	138.7
Dob-441 <1µm	illite, kaol., chl.	0.51		Lower anchizone	95		5			1086.5 ± 25.1	6.10	99.4	157.7
Dob-441 <0.5µm	illite, kaol., chl.	0.42		Upper anchizone						1063.3 ± 24.4	6.05		
Dob-449.1 <2µm	illite, kaol., chl.	0.29		Epizone	100			1033 ± 25		949.1 ± 22.2	6.40	99.5	138.6
Dob-449.1 2-0.5µm	illite, kaol., chl.	0.39		Upper anchizone						924.9 ± 21.4	6.63	99.8	138.9
Dob-449.1 1-0.5µm	illite, kaol., chl.	0.35		Upper anchizone						903.1 ± 20.9	6.73	99.6	136.8
Dob-449.1 <0.5µm	illite, kaol., chl.	0.37		Upper anchizone						912.6 ± 21.1	5.43	100	111.9
Dob-449.3 >2µm	kaol., illite, chl.	0.14		Epizone						1259.0 ± 29.1	0.402	100	12.70
Dob-449.3 <2µm	illite, kaol., chl.	0.19		Epizone									
Dob-449.3 2-1µm	kaol., chl., illite	0.21		Epizone						1047.7 ± 24.2	0.294	100	7.24
Dob-449.3 1-0.5µm	kaol., chl., illite	0.27		Epizone						1117.2 ± 25.8	0.554	99.4	14.87
Dob-449.3 0.5-0.2µm	kaol., chl., illite	0.25		Epizone						950.9 ± 22.0	6.02	99.7	130.7
Dob-449.3 <0.5µm	kaol., chl., illite	0.25		Epizone						1004.4 ± 23.2	0.428	99.8	9.98
Dob-476.6 >2µm	illite, kaol., chl.	0.26		Epizone						1170.4 ± 27.4	2.15	99.5	61.43
Dob-476.6 2-1µm	illite, kaol., chl.	0.29		Epizone	100					975.7 ± 22.2	4.43	99.4	101.7
Dob-476.6 <2µm	illite, kaol., chl.	0.33		Upper anchizone	100			1000 ± 12	994.6 ± 2.2	983.7 ± 23.0	4.80	99.4	108.9
Dob-476.6 <1µm	illite, kaol., chl.	0.31		Epizone	100					922.2 ± 21.2	6.18	100.0	129.0
Host whole rock													
JC-360.6 <2µm	illite	0.48		Lower anchizone									
JC-360.6 2-1µm	illite	0.68		Diagenetic						1066.9 ± 24.6	6.13	99.61	157.7
JC-360.6 1-0.5µm	illite	0.82		Diagenetic						1065.4 ± 24.6	6.90	99.7	173.8
JC-360.6 0.5-0.2µm	illite	0.52		Lower anchizone						1053.8 ± 24.3	6.68	99.2	165.8
JC-360.6 <0.2µm	illite	0.78		Diagenetic						928.3 ± 47.6	6.63	99.05	139.6
JC-490.2 <2µm	illite, chl.	0.46	0.36	Lower anchizone									
JC-500 <2µm	illite, chl.	0.51	0.36	Lower anchizone									
JC-500 2-1µm	illite, chl.	0.66		Diagenetic						1092.0 ± 25.2	3.37	99.69	87.71
JC-500 1-0.5µm	illite, chl.	0.62		Diagenetic						1076.8 ± 24.8	5.15	99.67	131.6
JC-500 0.5-0.2µm	illite, chl.	0.64		Diagenetic						1066.7 ± 24.5	5.39	99.47	135
JC-500 <0.2µm	illite, chl.	0.63		Diagenetic						878.3 ± 45.1	5.41	99.02	106.2
Dob-499.4 >2µm	illite, kaol., chl.	0.22		Epizone						1156.2 ± 26.6	0.37	97.1	10.26
Dob-499.4 <2µm	illite, kaol., chl.	0.20		Epizone									
Dob-499.4 2-1µm	illite, kaol., chl.	0.23		Epizone						1115.8 ± 25.7	2.03	99.6	54.38
Dob-499.4 1-0.5µm	illite, kaol., chl.	0.20		Epizone						1047.3 ± 24.1	5.08	99.5	125.1
Dob-499.4 0.5-0.2µm	illite, kaol., chl.	0.18		Epizone						1068.2 ± 24.6	3.30	99.3	83.41
Dob-499.4 <0.2µm	illite, kaol., chl.	0.21		Epizone						1025.6 ± 52.6	3.63	97.5	86.97

Calibration of the illite crystallinity (IC) and chlorite crystallinity (ChC) values and determination of very-lowe-grade- metamorthic zones has been done according to Warr and Mählmann (2015) and Warr and Cox (2016). kaol. = kaolinite, chl. = chlorite.

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STANDARD ID	K [%]	Rad. ^{40}Ar (mol/g)	Rad. ^{40}Ar (%)	Age (Ma)
HD-B1-137	7.96	3.3431E-10	89.76	24.06
LP6-151	8.37	1.9477E-09	97.28	129.41
HD-B1-139	7.96	3.4214E-10	92.33	24.63
LP6-153	8.37	1.9465E-09	97.32	129.33
HD-B1-140	7.96	3.3805E-10	92.39	24.33
LP6-154	8.37	1.9196E-09	97.39	127.61
HD-B1-141	7.96	3.4399E-10	92.86	24.76
LP6-155	8.37	1.9304E-09	97.69	128.30
HD-B1-142	7.96	3.4500E-10	93.28	24.83
LP6-156	8.37	1.9285E-09	97.59	128.17
HD-B1-147	7.96	3.4124E-10	92.88	24.56
LP6-161	8.37	1.9257E-09	97.13	127.99
HD-B1-148	7.96	3.3633E-10	90.67	24.21
LP6-162	8.37	1.9236E-09	97.21	127.80
HD-B1-149	7.96	3.3562E-10	90.93	24.16
LP6-163	8.37	1.9234E-09	97.21	127.85

Airshot ID	^{40}Ar/^{36}Ar	+/-
AS131-AirS-1	295.67	0.45
AS131-AirS-2	293.43	0.46
AS133-AirS-1	298.42	0.14
AS133-AirS-2	297.35	0.29
AS134-AirS-1	295.81	0.14
AS134-AirS-2	296.65	0.08
AS135-AirS-1	296.59	0.13
AS135-AirS-2	296.76	0.17
AS136-AirS-1	295.22	0.24
AS136-AirS-2	296.69	0.27
AS141-AirS-1	295.85	0.28
AS141-AirS-2	296.53	0.23
AS142-AirS-1	294.87	0.23
AS142-AirS-2	296.52	0.18
AS143-AirS-1	294.16	0.20
AS143-AirS-2	296.81	0.17

HD-B1: Hess and Lippolt (1994).

LP-6: Odin et al. (1982).
Recommended ^{40}Ar/^{36}Ar value: 295.5: Steiger and. Jäger (1977).

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Table 3. ^{87}Rb-^{86}Sr data for the untreated (U) and acid leached residues (R) of different clay fractions from the Milungera Basin fault gouges.

Sample	Size farction (µm)	Rb (ppm)	Sr (ppm)	^{87}Rb/^{86}Sr	^{87}Sr/^{86}Sr	±2σ
JC-343.3U	<2	263.8	244.7	3.14	0.771035	0.000008
JC-360.7U	<2	270.5	303.9	2.59	0.757396	0.000009
JC-360.7R	<2	253.7	269.0	2.74	0.759576	0.000009
JC-360.7R	2-1	279	361.1	2.25	0.754344	0.000008
JC-360.7L	2-1	61.96	914.2	0.678	0.715331	0.000006
JC-360.7R	<0.5	288.7	222.1	1.30	0.778401	0.000005
JC-360.7L	<0.5	68.94	906.2	0.076	0.715632	0.000006
JC-387.8U	<2	203.7	564.6	1.05	0.734125	0.000009
JC-387.8R	<2	201.6	537.4	1.09	0.735621	0.000007
JC-408U	<2	273.0	259.4	3.06	0.768788	0.000009
JC-440.5U	<2	206.8	637.0	0.941	0.733108	0.000007
JC-440.5R	<2	213.2	627.4	0.986	0.734396	0.000009
JC-440.5R	<0.5	279.2	240.1	1.16	0.765505	0.000006
JC-440.5L	<0.5	26.75	378.4	0.071	0.716576	0.000006
Dob-389.6U	2	239.5	184.5	3.78	0.778229	0.000009
Dob-389.6R	2-1	255.0	254.5	1.00	0.769394	0.000006
Dob-389.6R	<0.5	268.6	201.0	1.34	0.783259	0.000008
Dob-389.6L	<0.5	22.08	612.8	0.04	0.712000	0.000006
Dob-441U	<2	134.8	121.6	3.23	0.770115	0.000008
Dob-441R	2-1	200.7	203.4	0.987	0.767702	0.000006
Dob-441R	<1	202.1	201.8	1.00	0.768636	0.000007
Dob-441L	<1	20.5	399.5	0.051	0.720827	0.000006
Dob-449.1U	<2	214.7	337.0	1.85	0.750186	0.000009
Dob-449.1R	<2	218.9	326.1	1.95	0.751971	0.000014
Dob-449.1R	2-1	242.0	378.8	0.639	0.749952	0.000006
Dob-449.1L	2-1	13.59	265.7	0.051	0.718102	0.000006
Dob-476.6U	<2	190.0	521.6	1.06	0.744176	0.000009
Dob-476.6R	<2	200.2	517.7	1.12	0.744209	0.000007

U= untreated, R = residue, L = leachate

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Table 4. Trace element data (ppm) for the fault gouge illites.

Sample	JC-360.7	JC-387.8	JC-440.5	Dob-389.6	Dob-449.1	Dob-476.6	PAAS
La	89.8	281.5	242.3		328.0	396.2	38.0
Ce	193.9	619.1	519.4		591.8	773.3	80.0
Pr	19.9	63.1	54.3		73.1	80.2	8.90
Nd	69.1	225.6	187.3		257.0	271.5	32.0
Sm	12.8	39.1	32.2		44.5	46.3	5.60
Eu	2.22	4.32	4.22		7.30	7.01	1.10
Gd	9.27	24.3	21.2		29.8	28.7	4.70
Tb	1.05	2.07	2.59		3.70	3.23	0.77
Dy	5.09	7.22	12.9		20.2	15.9	4.40
Ho	0.94	1.17	2.42		4.36	3.29	1.00
Er	2.59	3.27	6.66		14.2	10.5	2.90
Tm	0.37	0.43	0.97		2.47	1.74	0.40
Yb	2.36	2.81	6.29		18.0	12.3	2.80
Lu	0.35	0.41	0.93		2.87	1.93	0.43
Th	80.8	110.9	83.6		107.7	149.1	14.60
U	7.54	8.43	18.6		32.2	40.3	3.10
Sc	20.7	10.2	11.2		91.4	58.1	16.00
$(La/Lu)_c$	29	23	76		13	29	10
Th/Sc	3.9	10.9	7.5		1.2	2.6	0.9
Th/U	10.7	13.2	4.5		3.3	3.7	4.7

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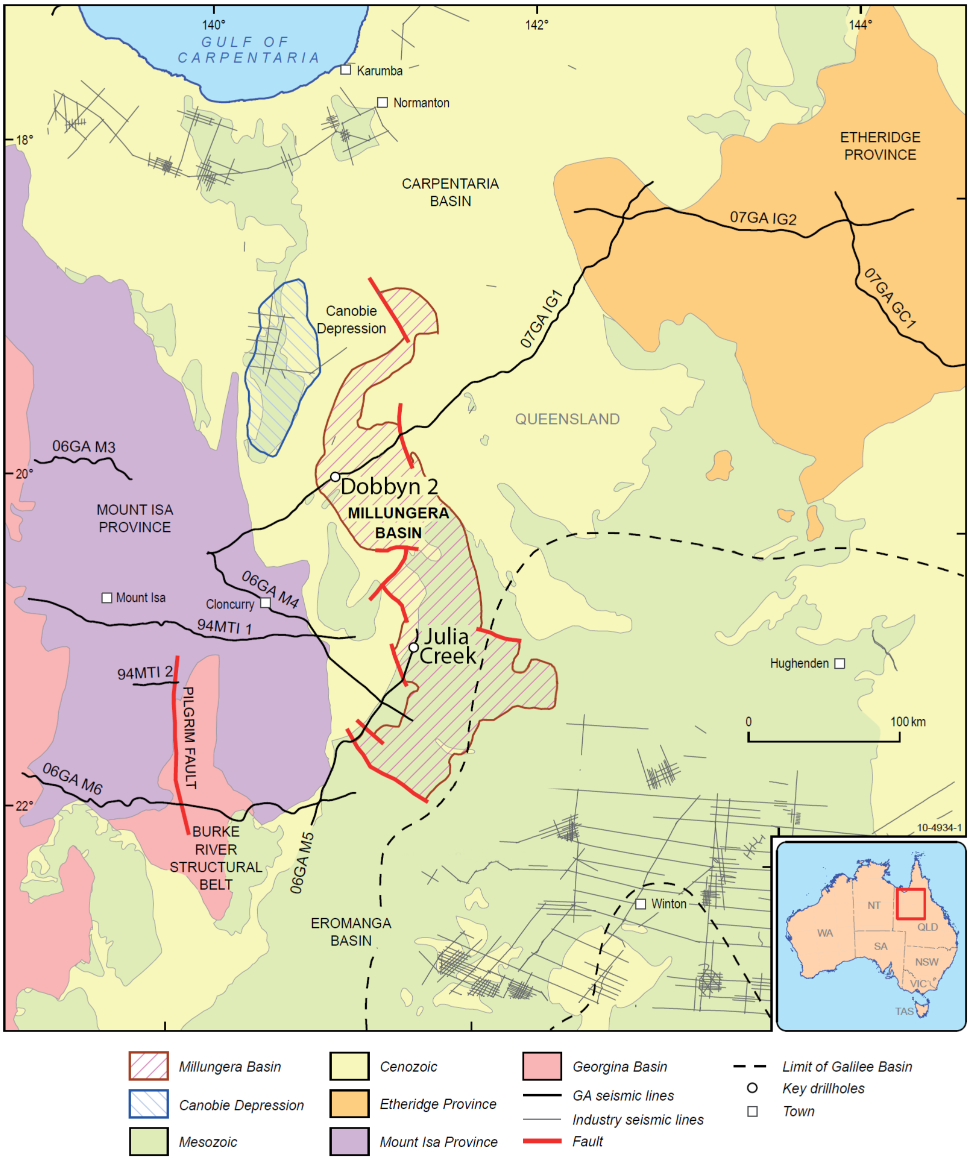

Figure 1

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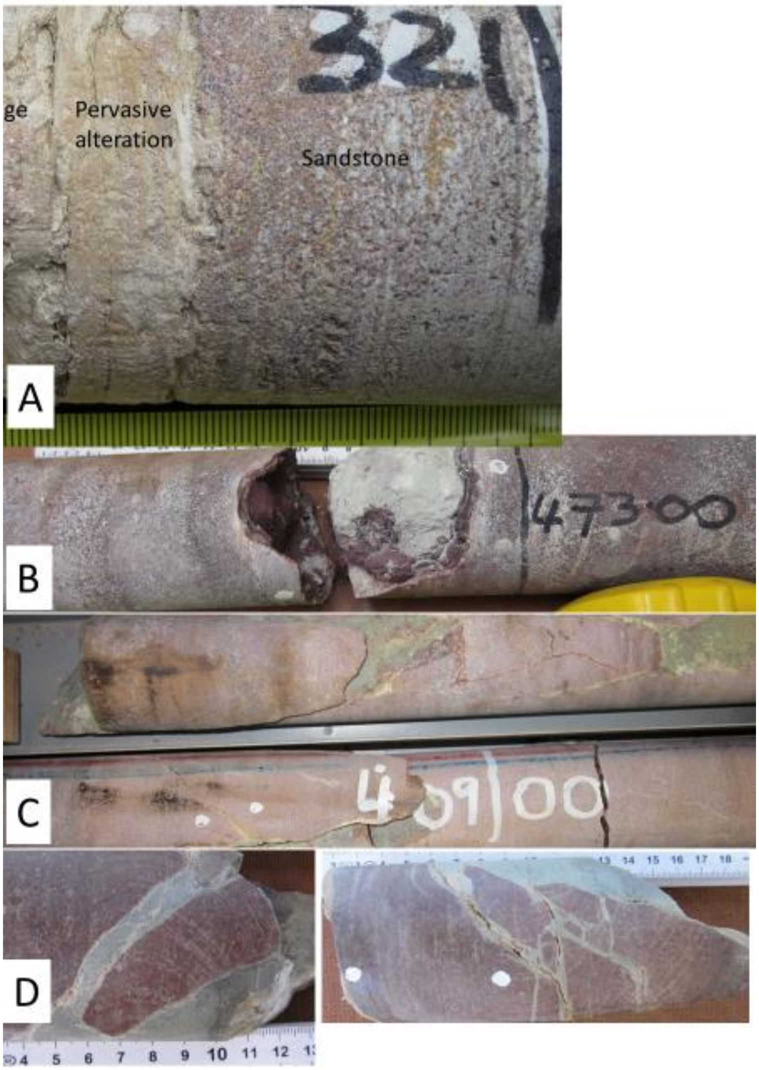
998 

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999 Fig. 2

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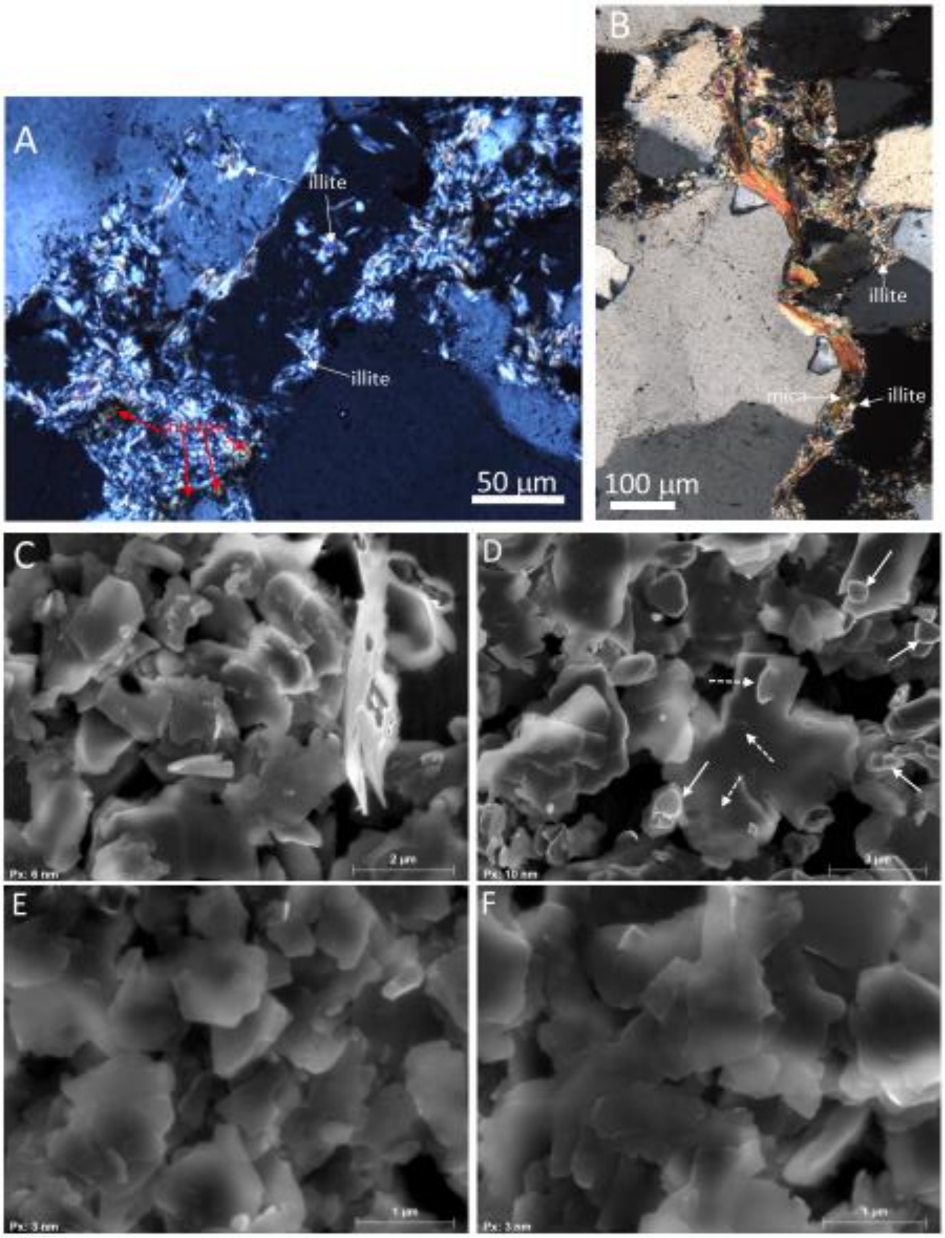

Figure 3.

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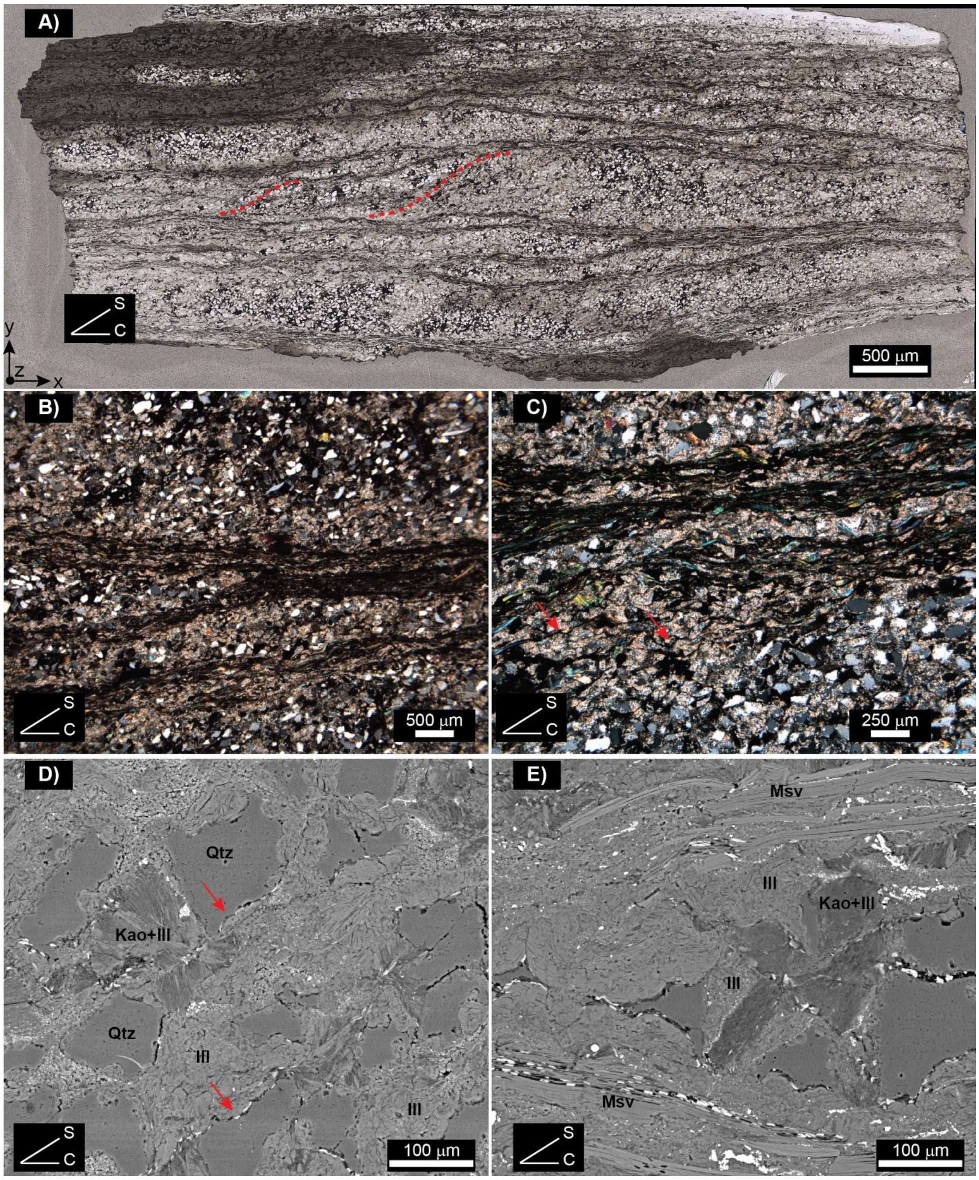

1002 Figure 4

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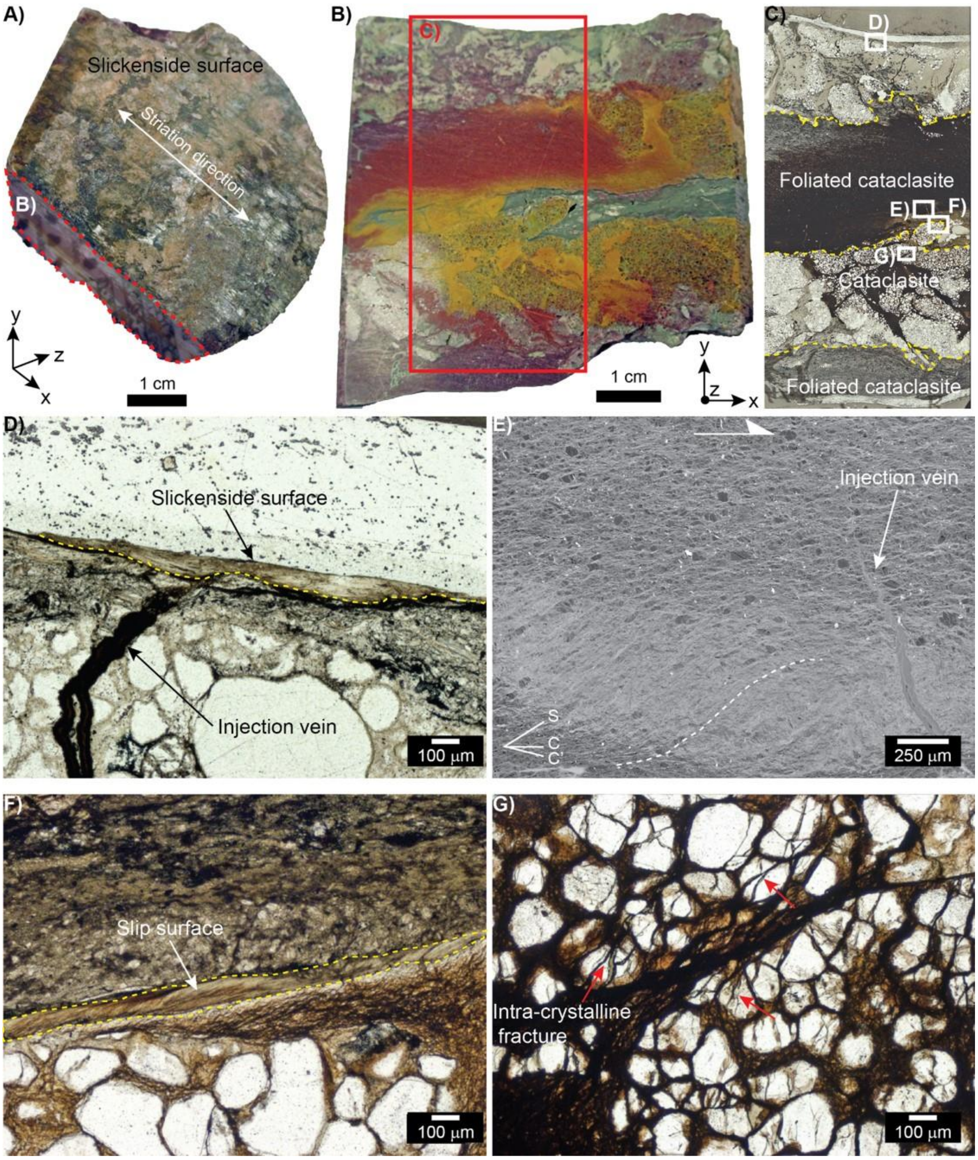

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1004 Figure 5

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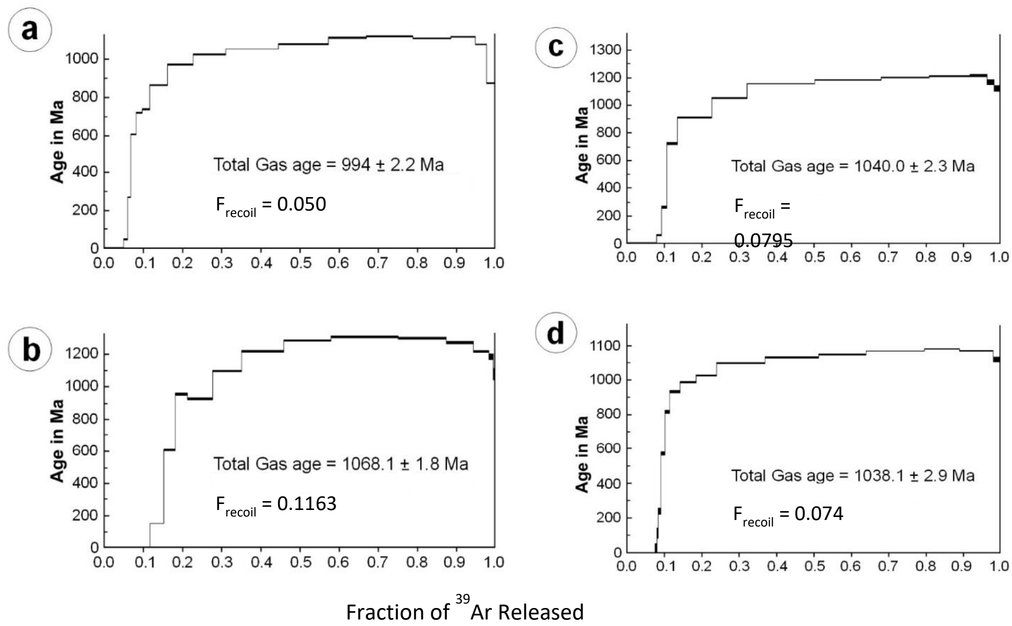

Figure 6

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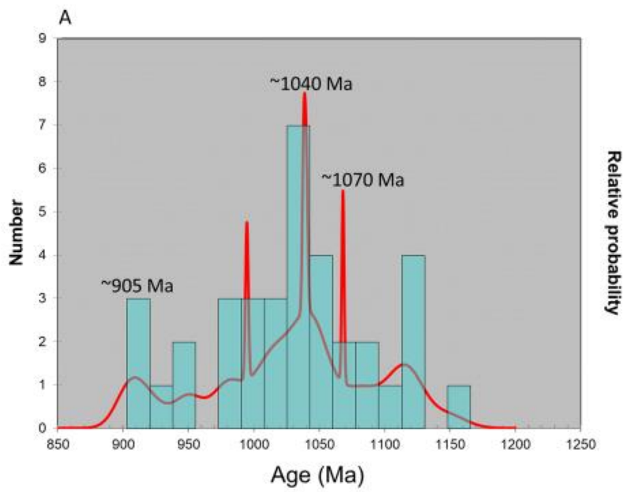

Figure 7a

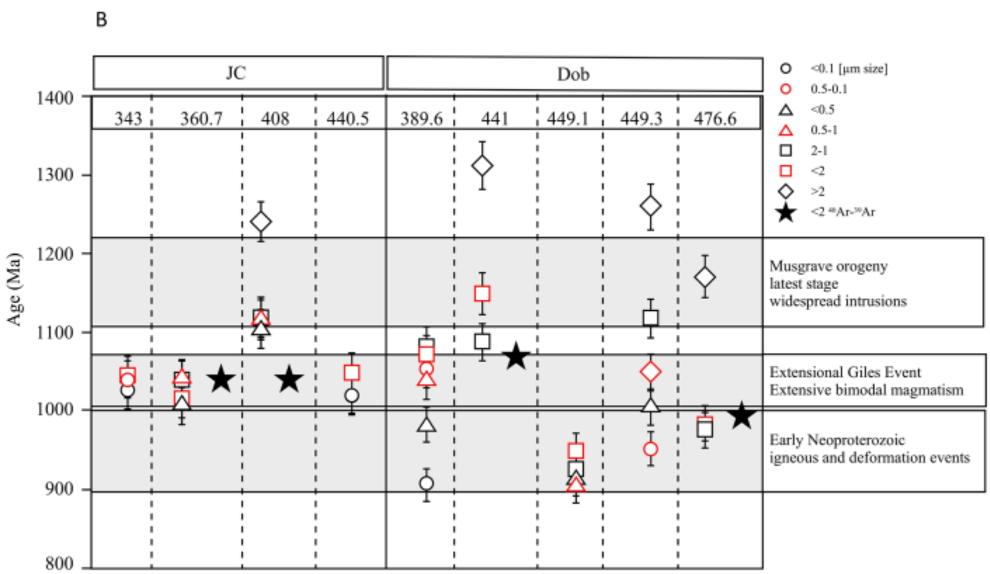

Figure 7b

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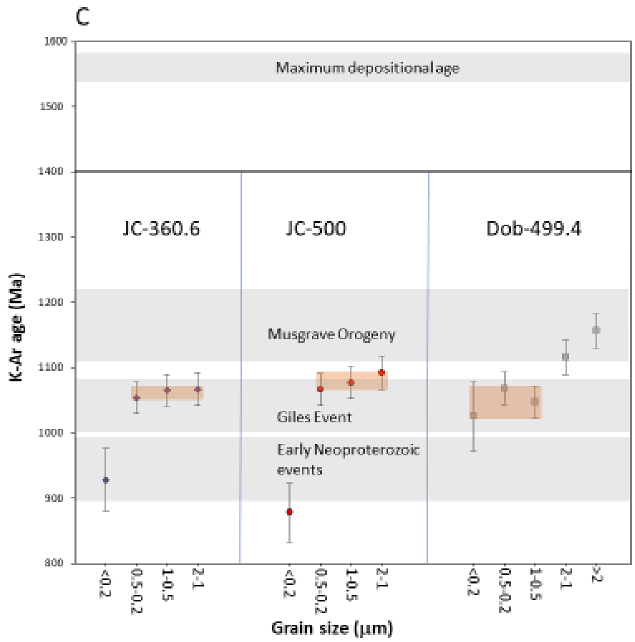

Figure 7c

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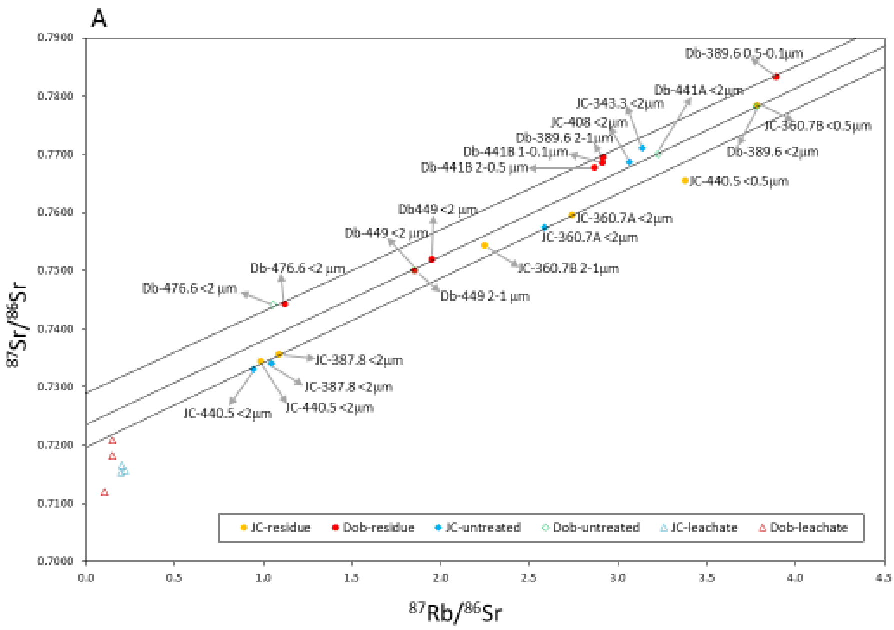

Figure 8a

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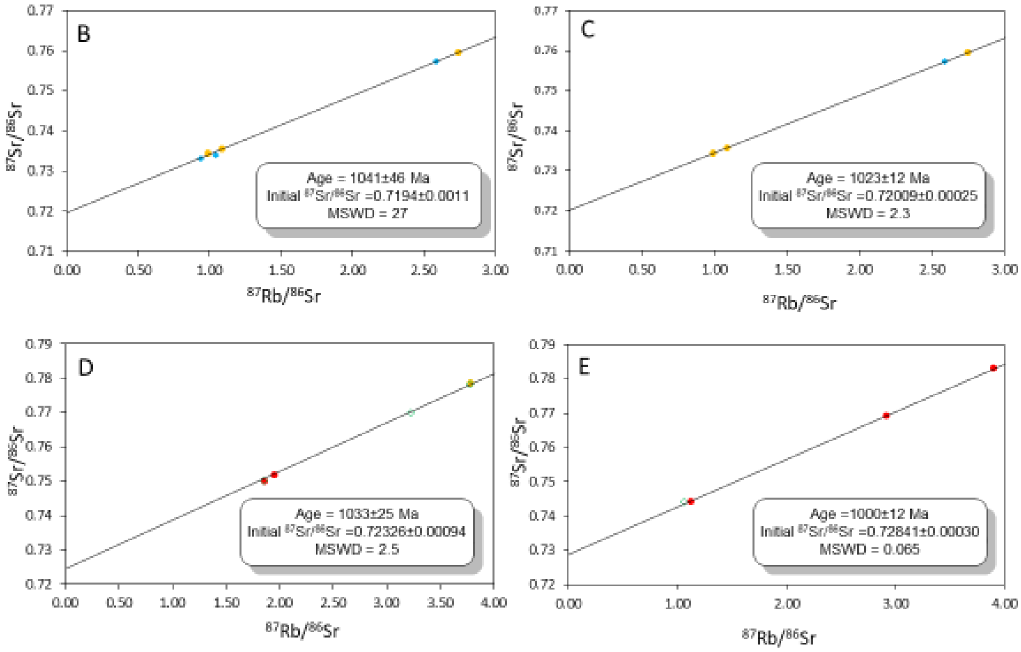

Figure 8b-e

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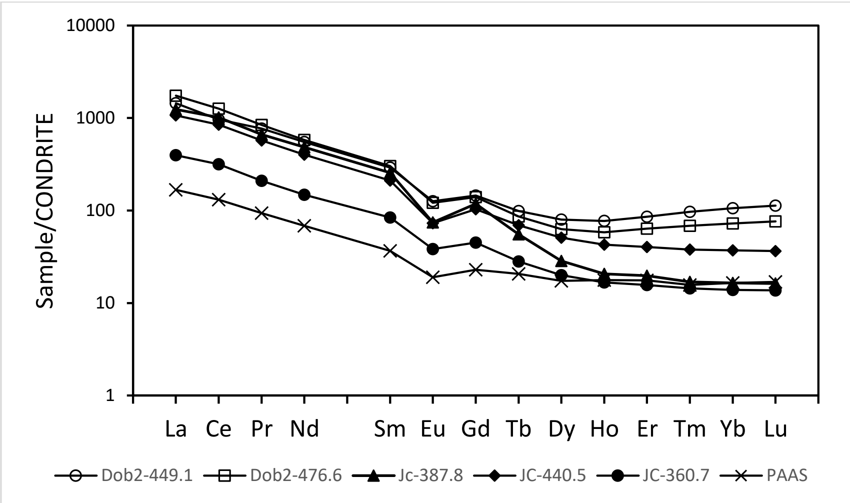

1026 Figure 9

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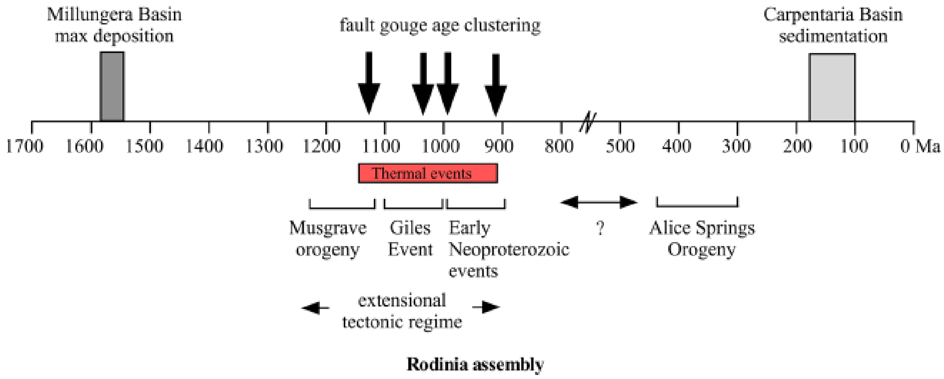

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1029 Figure 10

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