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Re-evaluation of the petrogenesis of the Proterozoic Jabiluka unconformity-related uranium deposit, Northern Territory, Australia

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Abstract The world class Jabiluka unconformity-related uranium deposit in the Alligator Rivers Uranium Field, Australia, contains >163,000 tons of contained U_3O_8 . Mineralization is hosted by shallow-to-steeply dipping basement rocks comprising graphitic units of chlorite–biotite–muscovite schist. These rocks are overlain by flat-lying coarse-grained sandstones belonging to the Kombolgie Subgroup. The deposit was discovered in 1971, but has never been mined. The construction of an 1,150 m decline into the upper eastern sector of the Jabiluka II deposit combined with closely spaced underground drilling in 1998 and 1999 allowed mapping and sampling from underground for the first time. Structural mapping, drill core logging and petrographic studies on polished thin sections established a detailed paragenesis that provided the framework for subsequent electron microprobe and X-ray diffraction, fluid inclusion, and O–H, U–Pb and $^{40}Ar/^{39}Ar$ isotope analysis. Uranium mineralization is structurally controlled within semi-brittle shears that are sub-conformable to the basement stratigraphy, and breccias that are developed within the hinge zone of fault-related folds adjacent to the shears. Uraninite is intimately associated with chlorite, sericite, hematite $\pm$ quartz. Electron microprobe and X-ray diffraction analysis of syn-ore illite and chlorite indicates a mineralization temperature of 200°C. Pre- and syn-ore minerals extracted from the Kombolgie Subgroup overlying the deposit and syn-ore alteration minerals in the Cahill Formation have $\delta^{18}O_{fluid}$ and δD_{fluid} values of 4.0 ± 3.7 and $-27 \pm 17\text{‰}$, respectively. These values are indistinguishable from illite separates extracted from diagenetic aquifers in the Kombolgie

Subgroup up to 70 km to the south and east of the deposit and believed to be the source of the uraniferous fluid. New fluid inclusion microthermometry data reveal that the mineralising brine was saline, but not saturated. U–Pb and $^{207}Pb/^{206}Pb$ ratios of uraninite by laser-ablation ICP-MS suggest that massive uraninite first precipitated at ca. 1,680 Ma, which is coincident with the timing of brine migration out from the Kombolgie Subgroup as indicated by $^{40}Ar/^{39}Ar$ ages of $1,683 \pm 11$ Ma from sandstone-hosted illite. Unmineralized breccias cemented by chlorite, quartz and sericite cross-cut the mineralized breccias and are in turn cut by straight-sided, high-angle veins of drusy quartz, sulphide and dolomite. U–Pb and $^{207}Pb/^{206}Pb$ ratios combined with fluid inclusion and stable isotope data indicate that these post-ore minerals formed when mixing between two fluids occurred sometime between ca. 1,450 and 550 Ma. Distinct $^{207}Pb/^{206}Pb$ age populations occur at ca. $1,302 \pm 37$, $1,191 \pm 27$ and 802 ± 57 Ma, which respectively correlate with the intrusion of the Maningkorri/Mudginberri phonolitic dykes and the Derim Derim Dolerite between 1,370 and 1,316 Ma, the amalgamation of Australia and Laurentia during the Grenville Orogen at ca. 1,140 Ma, and the break-up of Rodinia between 1,000 and 750 Ma.

Introduction

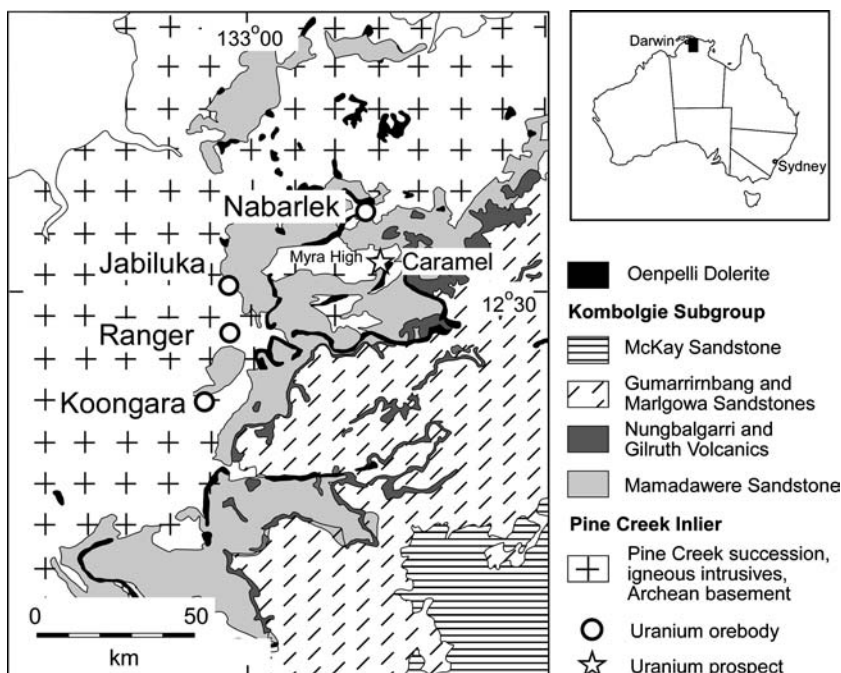
The Jabiluka unconformity-related uranium deposit is located in the Alligator Rivers Uranium Field within the Pine Creek geosyncline, approximately 225 km east of Darwin and 20 km north of the township of Jabiru in the Northern Territory, Australia (Fig. 1). It comprises two areas of mineralization, Jabiluka I and Jabiluka II (Fig. 2). Mineralization is hosted in a sequence of graphitic chlorite–biotite–muscovite schists belonging to the Paleoproterozoic Cahill Formation. Jabiluka I has a resource of 1.3 Mt of ore grading 0.25% U_3O_8 , whereas Jabiluka II is a world-class mineral resource of 31.1 Mt

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Fig. 1 Regional Geology of the east Alligator Rivers Uranium Field (ARUF) showing the location of Jabiluka in relation to other major deposits in the area. Modified after Needham (1988)

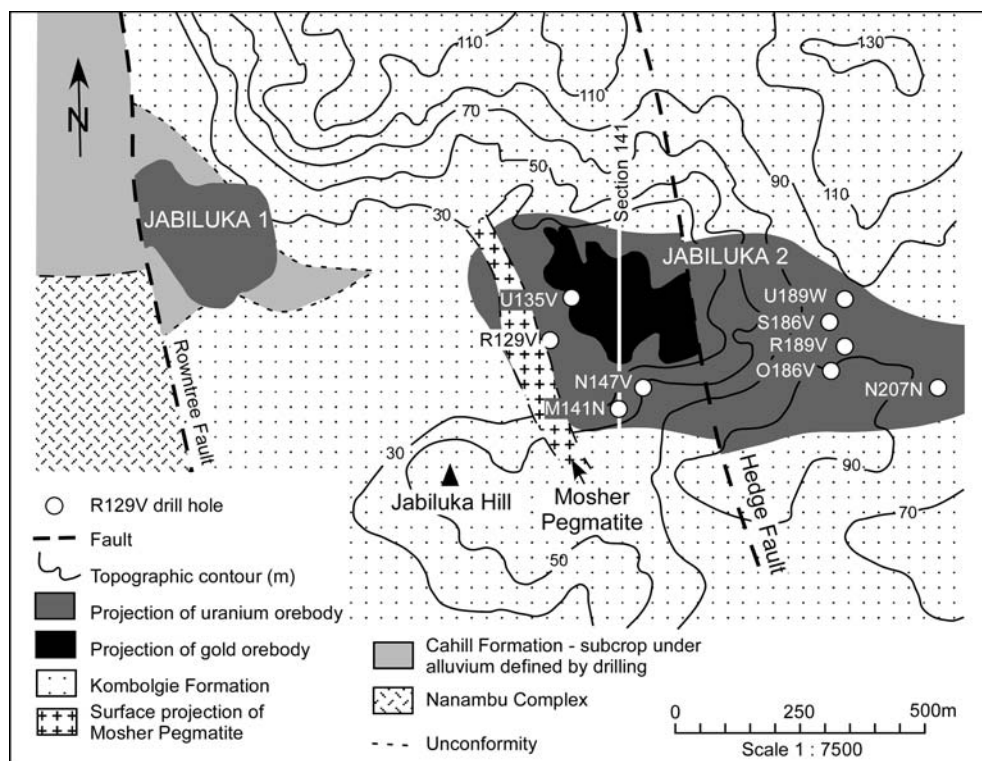


grading 0.53% U_3O_8 for 163,000 t of contained U_3O_8 . Jabiluka II also contains a significant gold resource that is estimated to be 4.6 Mt at a grade of 3.07 gpt Au.

Numerous publications document the geology of the Jabiluka uranium deposit (Rowntree and Mosher 1975; Hegge 1977; Gustafson and Curtis 1983; D. J. Johnston, unpublished PhD thesis; Needham 1988; Hancock et al.

1990) and much of the work presented in this paper builds on these results. There are also many papers that describe geochemical aspects of the deposit, but there are still many unresolved questions about the origin and timing of the ore fluids. For example, Binns et al. (1980a), Ewers and Ferguson (1980) and Nutt (1989) describe the petrology and chemical composition of the

Fig. 2 Plan map showing the geology and surface topography of the Jabiluka uranium deposit with Jabiluka I and II and the Mosher pegmatite projected through the cover of the Kombolgie Subgroup (modified after Ewers et al. 1984). Drill holes used in this study and section 141 represented in Fig. 4 are shown



various chlorite phases at Jabiluka assuming that all are coeval with uraninite, whereas it is known that there are pre-, syn- and post-ore generations of chlorite at Jabiluka. Oxygen, carbon and sulphur stable isotopic data from quartz, sulphide and carbonate phases have been reported by Binns et al. (1980b) and Ypma and Fuzikawa (1980), but many of the minerals analysed in these studies may be post-ore. Timing constraints on basin formation and ore-precipitation have been the subject of at least four papers (Hills and Richards 1976; Page et al. 1980; Ludwig et al. 1987; Mass 1989). However, the recently determined single-grain U–Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ ages from a variety of intrusive and extrusive igneous rocks in northern Australia (Jagodzinski 1998; Carson et al. 1999; Sweet et al. 1999a, b; Rawlings 1999; Kyser et al. 2000; Edgecombe et al. 2002) have shed new light on the geochronology understanding of the region and the district. Finally, fluid inclusion data from quartz and carbonate veins collected at Jabiluka have been reported by Ypma and Fuzikawa (1980), Wilde et al. (1989) and Derome et al. (2003a), and these have been reinterpreted and augmented with new fluid inclusions, in light of the detailed paragenesis reported in the present paper.

The construction of an 1,150 m access decline in 1998 and 1999, followed by the first underground drilling program of the upper eastern sector of Jabiluka II in 2000 (Fig. 3) provided an opportunity to map and sample parts of the deposit that were previously inaccessible. Samples from historical and new drill core were also collected. The data gathered from this study was used to construct a detailed paragenesis that has been used to reinterpret the genesis of Jabiluka and other

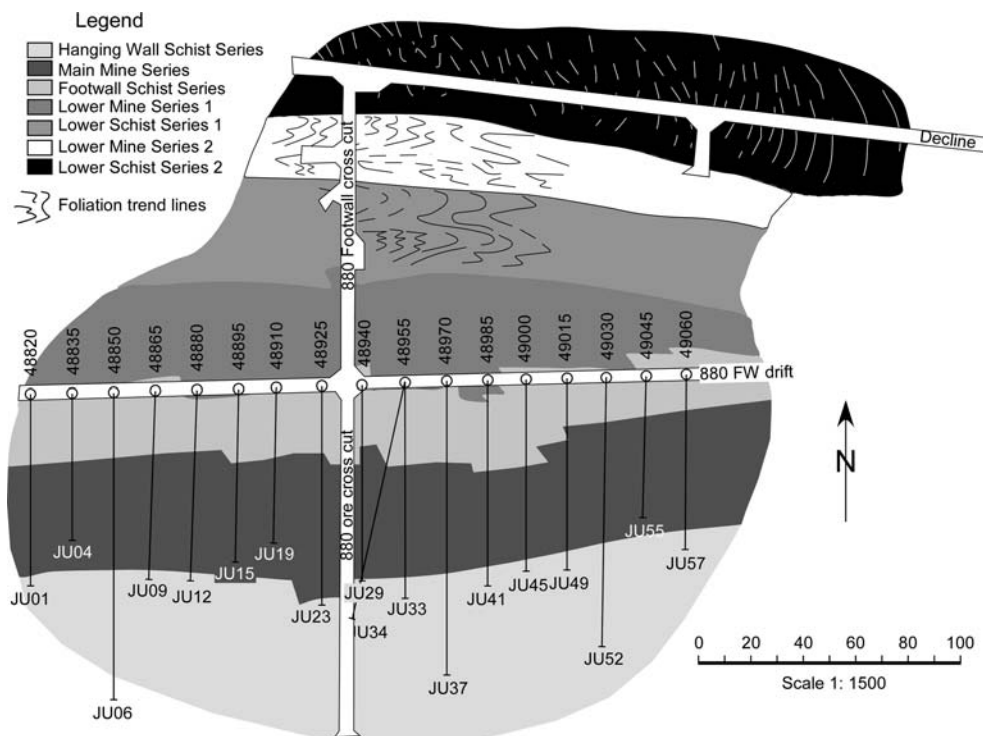
similar unconformity-type uranium deposits in northern Australia likely formed.

Geological considerations

The Jabiluka uranium deposit occurs within the Alligator Rivers uranium field (ARUF) along the northwest margin of the McArthur Basin. The geology of the ARUF is well documented (Ferguson 1980; Needham and Stuart-Smith 1980; Page et al. 1980; Ewers et al. 1984; Needham 1988) and only a brief summary is presented here. The basement complex of Archean granitoid gneisses and Paleoproterozoic metasedimentary rocks underwent prograde metamorphism during the Nimbuwa event of the top end orogeny (1,870–1,855 Ma), reaching lower to middle amphibolite facies in the area near the Jabiluka deposit. The Nimbuwa event was followed by post-peak regional retrogression that ended prior to ca. 1,820 Ma based on U/Pb ages of $1,828 \pm 5$ and $1,822 \pm 5$ Ma from the Pul Pul Rhyolite and the Plum Tree Creek Volcanics that were deposited on the basement rocks in the South Alligator Rivers Uranium Field and the ARUF, respectively (Jagodzinski 1998; Sweet et al. 1999b).

Jabiluka comprises of two areas of mineralization, Jabiluka I and Jabiluka II. The small Jabiluka I resource is bounded on the western side by the Rowntree Fault and located 500 m west of the world-class Jabiluka II ore body. The Jabiluka II ore body is hosted by an east–west trending, south-dipping sequence of predominantly chlorite–biotite–muscovite schists that form the lower

Fig. 3 A generalised geological plan map showing location of the 880 Ore crosscut, the 880 Footwall crosscut, the 880 Footwall Drift, which incorporates the 880 Footwall Drive–West and East, the decline from surface and the drill hole centres along the footwall drives. The drill holes shown represent the projection of the longest drill hole at each centre. ERA geologists mapped the contacts shown along the 880 Ore X-cut



member of the Cahill Formation (Table 1). The ore body is divided into a western and eastern block by the north-trending, steeply west-dipping (85°) Hegge Fault (Fig. 2). The host sequence to the uranium mineralization is divided into nine lithostratigraphic units (Table 1). The Upper Graphite Series, the Main Mine Series, the Lower Mine Series 1 and the Lower Mine Series 2 are carbonaceous/graphitic-bearing members that collectively host 92% of the identified mineralization. The Main Mine Series alone contains 70% of the identified resource (Table 1). In contrast, the Upper Schist Series, the Hanging Wall Schist Series, the Footwall Schist Series, the Lower Footwall Schist Series and the Lower Footwall Schist Series 2 are non-carbonaceous quartz–muscovite and chlorite schists containing limited mineralization (Table 1). A zone of significant gold enrichment occurs in the Main Mine Series immediately west of the Hegge Fault and a smaller gold zone occurs in the Lower Schist Series 2 to the east of the fault. Coarse-grained pegmatite dykes up to 30 m wide and lesser basic dykes are scattered throughout the Jabiluka deposit (Hegge 1977; Hancock et al. 1990). The unmineralized Mosher pegmatite (Fig. 2) is pre-Kombolgie in age (Hancock et al. 1990), but some of the basic dykes intruded after mineralization (Page et al. 1980).

The basement complex is overlain by unmetamorphosed, shallowly dipping sandstones, conglomerates and volcanic rocks belonging to the Kombolgie Subgroup (Fig. 4). The Kombolgie Subgroup is the basal succession of the Proterozoic Katherine River Group (Needham 1988; Sweet et al. 1999a) and was deposited between 1,822 and 1,723 Ma, as defined by the underlying Plum Tree Creek Volcanics and the intruding Oenpelli Dolerite, respectively (Sweet et al. 1999a; Kyser et al. 2000; Edgecombe et al. 2002). The Kombolgie Subgroup is comprised, in ascending stratigraphic order, of alluvial fan to braided fluvial facies of the Mamadawerre Formation, the mafic Nungbargarri Volcanics, the proximal to distal braided fluvial facies, to upper shoreface and eolian facies of the Gumarrirnbang Formation, the basaltic Gilruth Volcanics, and the distal braided fluvial to upper shoreface Marlgowa Formation (Sweet et al. 1999b; Hiatt and Kyser 2002). Interfingering with the Marlgowa Formation are glauconitic and phosphatic marine upper shoreface facies of the McKay Formation (Hiatt and Kyser 2002). The Kombolgie Subgroup is known to be at least 1,350 m thick at the southern margins of the ARUF (Holk et al. 2003) and may have been as thick as 1,810 m in the eastern Mount Marumba area (Sweet et al. 1999b).

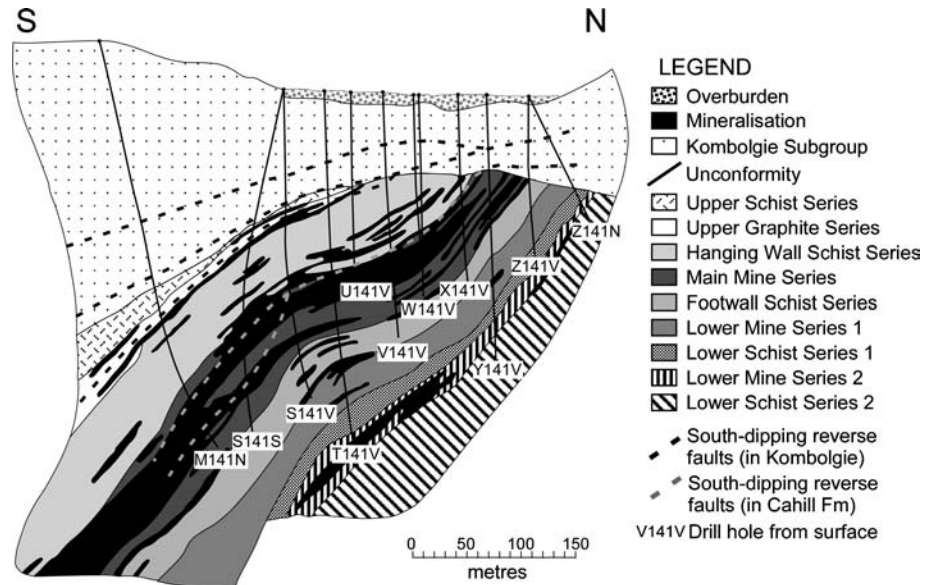
Geology at the unconformity

Structural measurements and macroscale observations compiled from underground drill holes and mapping on

Table 1 Stratigraphy of the Lower Cahill Formation, host of the Jabiluka deposit as shown in Fig. 3

Stratigraphic Unit and description
Upper Schist Series (2–13 m). Hosts 0% of known uranium mineralization Quartz–muscovite–chlorite and Muscovite–chlorite schists.
Upper Graphite Series (4–20 m). Hosts 4% of known U mineralization Fine-grained graphite schists and graphite–chlorite schists with quartz and pyrite lenses; fine-grained muscovite–chlorite schist with extremely fine-grained graphite and thin silicified dolomite/magnesite beds. Local breccias.
Hanging Wall Schist Series (60 m). Hosts 2% of known U mineralization Coarse- and fine-grained, muscovite-rich, chlorite–muscovite–sericite, chlorite–muscovite and chlorite–sericite schists with minor feldspar and biotite. Garnetiferous intervals, containing 5 to 15 % altered garnet porphyroblasts. Footwall contact with Main Mine Series is generally sharp and marked by the first appearance of carbonaceous material.
Main Mine Series (30–135 m). Hosts 70% of known U mineralization plus Au resource Fine- to medium-grained grey to black, carbonaceous ± pyritic chlorite schists; chlorite–sericite schists; chlorite breccias, thin chlorite–muscovite–feldspar schists and brecciated chert/jasper beds. Jasperoidal breccias may correlate with dolomite–magnesite beds in barren parts of the stratigraphy. Extensively brecciated. Muscovite-rich schists absent. Bleached schists border mineralised carbonaceous schists and are gradational into the surrounding units. These rocks are not separate stratigraphic units, but likely represent foliation / fracture-controlled alteration. Lower contact with the Footwall Schist Series is gradational over 2 m and marked by an increase in coarse-grained muscovite-rich layers.
Footwall Schist Series (5–65 m). Hosts 4% of known U mineralization Coarse- and fine-grained, muscovite-rich chlorite–muscovite, chlorite–sericite–muscovite, and muscovite–chlorite pelitic to semipelitic schists with minor feldspar and biotite. Contains abundant quartz segregation lenses or lenticles. Contact with the Lower Mine Series 1 is gradational over several metres
Lower Mine Series 1 (5–85 m). Hosts 6% of known U mineralization Sequence of thin, medium- to fine-grained variably carbonaceous to graphitic chlorite–sericite, quartz–muscovite and chlorite–sericite–biotite semipelitic schists with minor chloritic breccias at Jabiluka II.
Lower Footwall Schist Series 1 (5–15 m). Hosts 1% of known U mineralization Quartz–muscovite–chlorite and Chlorite–muscovite schists.
Lower Mine Series 2 (5–100 m). Hosts 12% of known U mineralization Sequence of medium- to fine-grained chlorite–graphite schists, chlorite–feldspar schists and a dolomite/magnesite bed up to 10 m thick. Minor chloritic breccias at Jabiluka II. Biotite–feldspar schists occur at Jabiluka I
Lower Footwall Series 2 (140 m). Hosts 1% of known U mineralization Chlorite–muscovite, quartz–chlorite–muscovite, chlorite–feldspar–muscovite and biotite–feldspar schists locally containing altered garnet porphyroblasts interbedded with fine- to medium-grained chloritised amphibolite. Hosts a small zone of gold enrichment east of the Hegge Fault

Fig. 4 Interpreted cross section through section 141 of the Jabiluka deposit (Fig. 2) showing the subhorizontal nature of the unconformity over the northern part of the deposit and the steeply south-dipping contact over the southern part of the deposit based on surface exploration drill hole data. The reverse fault structures indicated along the unconformity are those identified by D. J. Johnston (unpublished PhD Thesis)



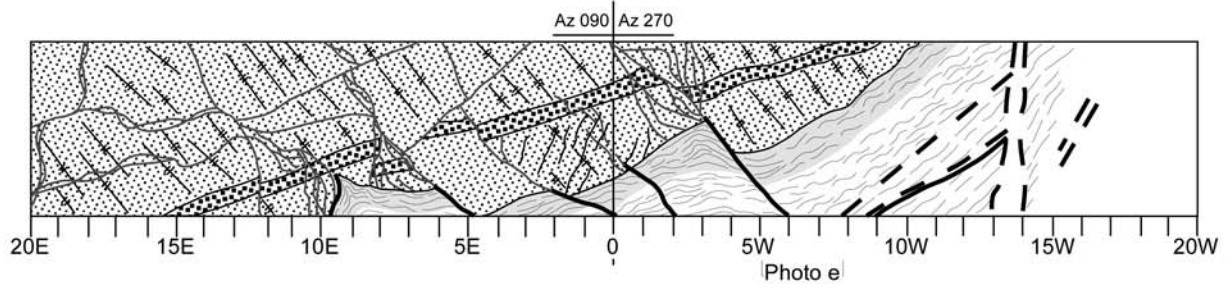
the 880 level (Fig. 3) are consistent with the four deformational events (D_1 – D_4) proposed by D. J. Johnston (unpublished PhD thesis), and provide new insights into pre- and post-ore deformation. The presence of the Kombolgie Subgroup has obscured the nature of the unconformity at Jabiluka and numerous theories have been proposed to explain why the sandstone–basement contact is largely subhorizontal over the northern part of the deposit, but steeply south-dipping with up to 300 m of relief over the southern part of the deposit (Fig. 4). Early workers proposed that the steep south-facing basement–sandstone contact was the South Fault (Rowntree and Mosher 1975; Hegge 1977; Hegge et al. 1980). Others proposed that a major period of karst development prior to the deposition of the Kombolgie Subgroup was responsible for the paleotopographic variation (Ferguson et al. 1980). Gustafson and Curtis (1983) later proposed that the basement–sandstone contact was a normal unconformity offset by minor faulting.

The unconformity exposed along the decline (Fig. 5) represents the basement–sandstone contact above the northern part of the deposit. The trace of the unconformity along the decline is irregular due to a series of cross-cutting, predominantly west-dipping normal listric faults that are consistent with an overall E–W extension (Fig. 5a, b). These faults rotate the basement–sandstone contact into a series of back-rotated blocks that offset the sandstone such that it is progressively down-dropped to the west and northwest, but gives the unconformity an overall dip to the east, consistent with the trend defined by surface drill hole intersections of the unconformity. The overall orientation of the listric faults is similar to the post-ore NNW-trending Rowntree and Hegge faults (Fig. 2) and other smaller NW- and NNW-trending cross faults and veins that offset mineralized reverse shear structures along the 880 underground development. Above the unconformity, the faults form a

bifurcating network of steeply-dipping chlorite-rich fractures and quartz–chlorite breccias (Fig. 5d). Vertical displacements of pebbly beds along these faults range from a few ten's of centimetres (Fig. 5e) to several metres and commonly expand into broad zones comprising subvertical anastomosing chlorite-lined fractures and breccias. These fractures vary from discrete narrow linear features, <1 cm wide, to more irregular curvilinear fractures and breccias, up to 50 cm wide. The fractures and breccias are filled with massive chlorite. An increase in chlorite in the breccia zones corresponds to a decrease in quartz grains. Here, wall rock contacts commonly exhibit subrounded shapes and vague diffuse outlines due to the progressive chlorite replacement (Fig. 5d). Below the unconformity, the listric faults typically flatten to merge with the basement foliation. A <1 m wide zone of weak to moderately hematitic pelitic schists is irregularly developed immediately below the unconformity (Fig. 5a). It is offset and pinched out by the listric faults suggesting that it was present prior to faulting. This hematite zone may represent a pre-Kombolgie paleoweathering remnant, or a sub-Kombolgie redox profile caused by the interaction of oxidised basal sandstone-hosted fluids with more reduced basement rocks. The listric faults may occur at a much larger-scale

Fig. 5 a, b The trace of the unconformity looking south (*above*) and north (*below*) along the decline above the northern part of the deposit. Note the highly irregular basement–sandstone contact due to the occurrence of cross-cutting, west-dipping normal listric faults. **c** An example of a closely spaced network of predominantly listric faults that occur in the sandstone immediately above the unconformity. **d, e** Close ups of the unconformity along the decline showing chlorite filled fractures <1 cm wide, evolving into irregular curvilinear fractures and breccias, up to 50 cm wide filled with massive chlorite. Note the subrounded sandstone shapes due to progressive chlorite replacement above the unconformity and the SiO_2 quartz boudin in the underlying basement schist

a Jabiluka Decline - Unconformity Section - South Panel (Looking South)



b Jabiluka Decline - Unconformity Section - North Panel (Looking North)

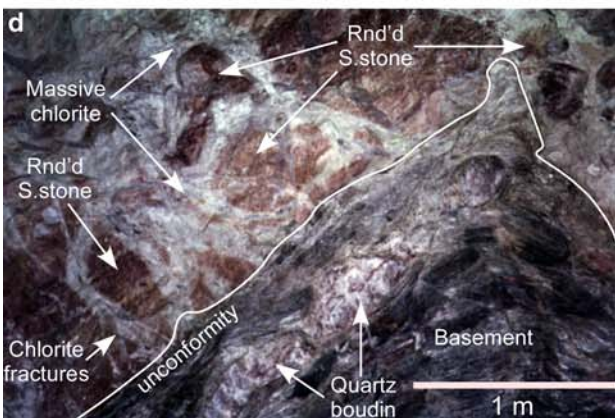
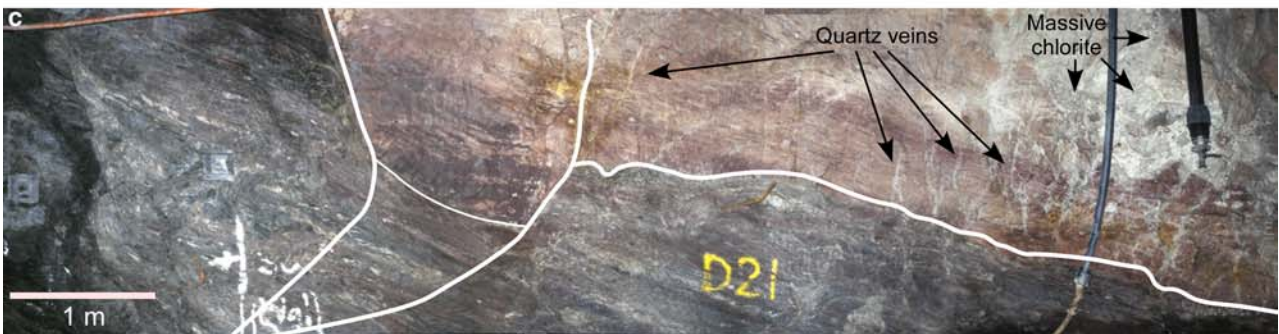
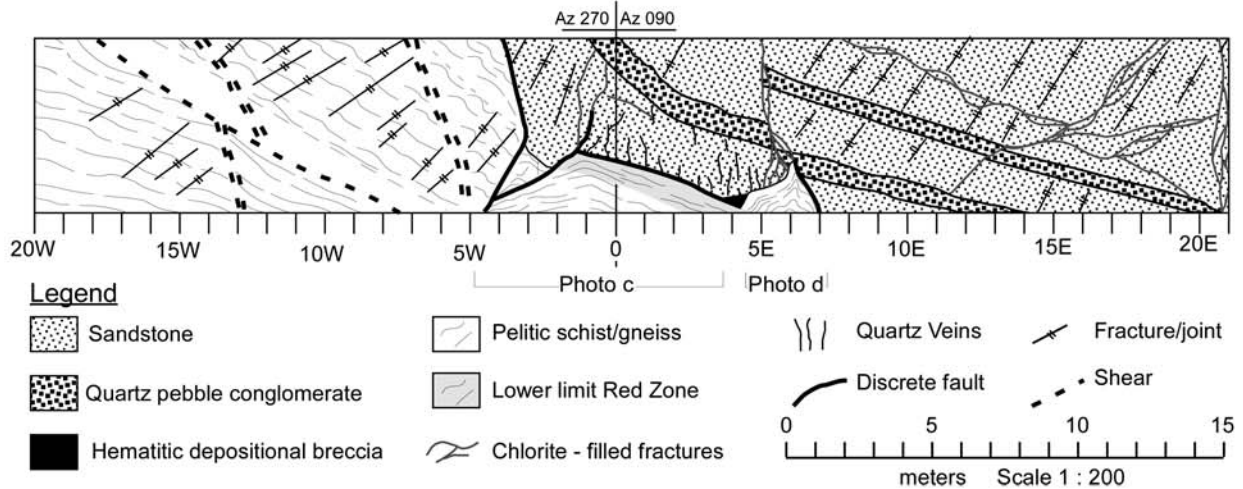
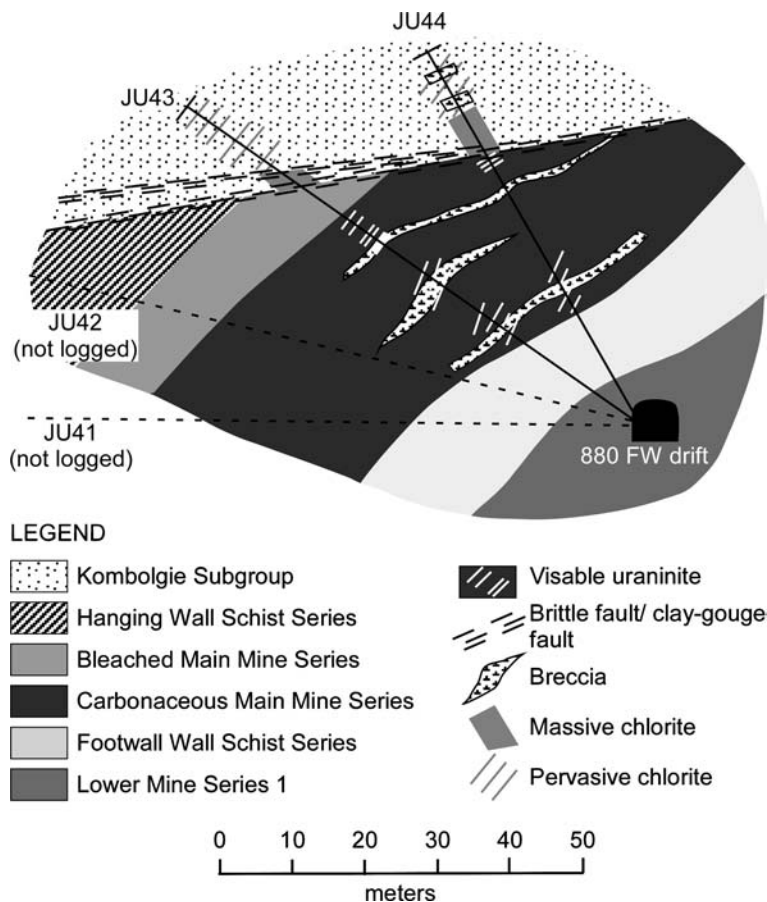


Fig. 6 Geological interpretation of a drill hole fan along section 48,985 m E (Fig. 3) suggesting that faulting along the basement-sandstone contact cuts the up-dip projection of zones containing visible uraninite supporting the observation that these structures are post-mineralization



and could off-set the unconformity several tens of metres or more, but are unlikely to account for the 300 m of steep, south-facing displacement observed over the southern side of the deposit (e.g. Fig. 4).

Logging of underground drill core identified a zone of brittle fault structures developed at, and immediately below, the basement-sandstone contact between sections 48,925 m E and 49,060 m E (Fig. 3). The structures comprise carbonaceous, clay-rich damage zones and localised brittle fault breccias and gouge-rich faults within dispersed damage zones. The carbonaceous, clay-rich zones range up to 1 m wide and contain poorly indurated, variably carbonaceous, clay-rich gouge that host sub-centimetre-scale angular fragments. Variably developed penetrative pressure-solution cleavage (P-fabric) and synthetic Reidel shear fractures typically bound these fragments. Damage zones vary up to 3 m wide and are distinguished by abundant cross-cutting chloritic slip surfaces and extensive fracture networks. The brittle fault breccias and gouge-rich faults include 3–4 m wide zones of chlorite-cemented basement-sandstone breccias or massive chlorite and narrow clay-rich gouges. In some drill holes (e.g. JU-32A), a sandstone wedge is structurally overlain by the basement schist that was thrust faulted into place. This basement wedge is in turn overlain by another basement-sandstone contact, which may be marked at its tops by a clay-rich, brittle-fault breccia zone. Based on 12 drill hole intersections

through the basement-sandstone contact, the fault zone is calculated to be a shallow, south-dipping structure oriented at $347^{\circ}/12^{\circ}$ and developed immediately below or at the unconformity in this area (Fig. 6). Kinematic evidence suggests that this subhorizontal structure had a reverse sense of displacement with a small sinistral component of horizontal movement. In some drill holes these structures cut-off the up-dip projection of zones containing visible uraninite (Fig. 6). In several cases, fragments or clasts of angular uraninite were noted within fault breccia intervals along the basement-sandstone contact, thereby suggesting that this horizontal structure is post-mineralized. These observations are consistent with those described by D. J. Johnston (unpublished PhD thesis) who noted that a reverse fault structure with a moderate dip to the south occurs along the unconformity and is related to multiple zones of brecciation and fault related foliation in the overlying Kombolgie Subgroup above the hinge of the south-dipping slope in the area he studied (Fig. 4). This, he concluded, was the product of east-west trending, south-dipping reverse faults that followed foliation and zones of weakness in the basement lithologies, the subhorizontal contact along the unconformity and which coalesced as low angle faults in the middle levels of Kombolgie Subgroup (Fig. 4). These reverse structures have the required orientation to explain the south-facing topography that occurs over the southern side of the

deposit, but they cannot explain the apparently normal sense of displacement that is evident.

Based on our observations, we conclude that the steep south-facing basement–sandstone contact with up to 300 m of displacement is a normal unconformity contact that may have been offset by minor faulting, concurrent with the findings of Gustafson and Curtis (1983). Although the topographic relief in this area may seem excessive given that highly fissile, mica schists are not particularly resistant in a normal weathering environment, Jabiluka sits on the south-western margin of the much larger Myra High (Fig. 1). This paleotopographic land-form is now known to have existed prior to the deposition of the Kombolgie Subgroup based on the sequence stratigraphic mapping and paleocurrent indicators (Hiatt and Kyser 2002) and possibly explains the presence of a topographic shoulder in the vicinity of the Jabiluka deposit. Underground mapping and logging of drill core suggest that the shallowly dipping reverse structures along the unconformity were formed prior to

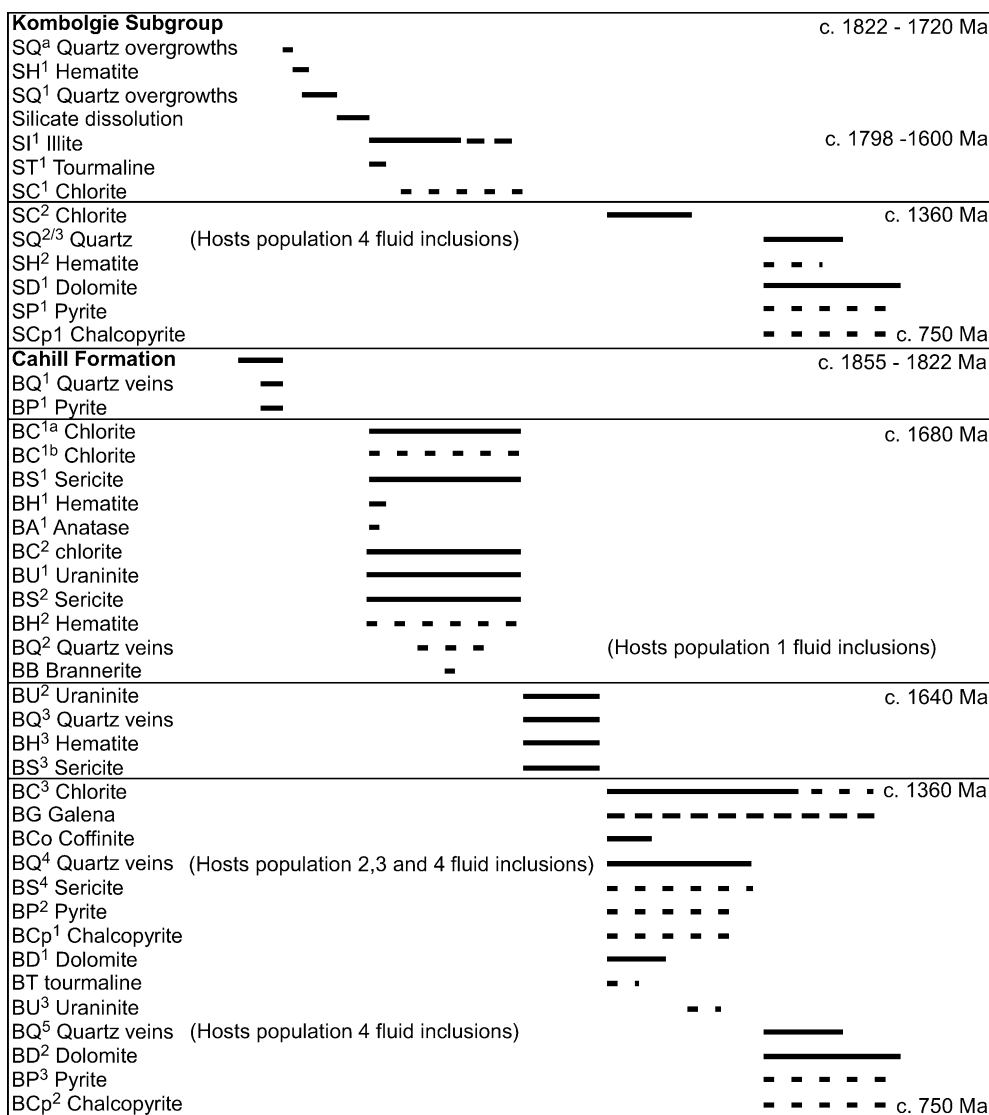
the NNW trending west-dipping normal listric faults and fractures that are filled with drusy quartz and chlorite.

Methods

More than 300 hand specimens were collected from the 880 m development level and the ERA exploration and core storage facility on the Jabiluka mining lease. Polished thin sections were prepared for all samples used for geochemical analysis. The thin sections were examined using transmitted and reflected light to determine the mineral paragenesis. Electron microprobe analyses were performed on polished thin sections using a Cambax MBX electron microprobe equipped with 4 WDX X-ray spectrometers at Carleton University, Ottawa, Canada.

Microthermometric measurements on doubly-polished thick sections were performed using a Linkam TH600 heating–cooling stage by standard techniques

Fig. 7 Paragenetic diagram summarizing the relative and absolute timing of alteration assemblages in the sandstone and basement lithologies around the Jabiluka deposit. The S and B before each phase denotes sandstone diagenesis/alteration and basement alteration respectively. The ages presented for each event are based on cross-cutting relationships that have been constrained by K/Ar ages (Page et al. 1980) and $^{40}\text{Ar}/^{39}\text{Ar}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages (this study)



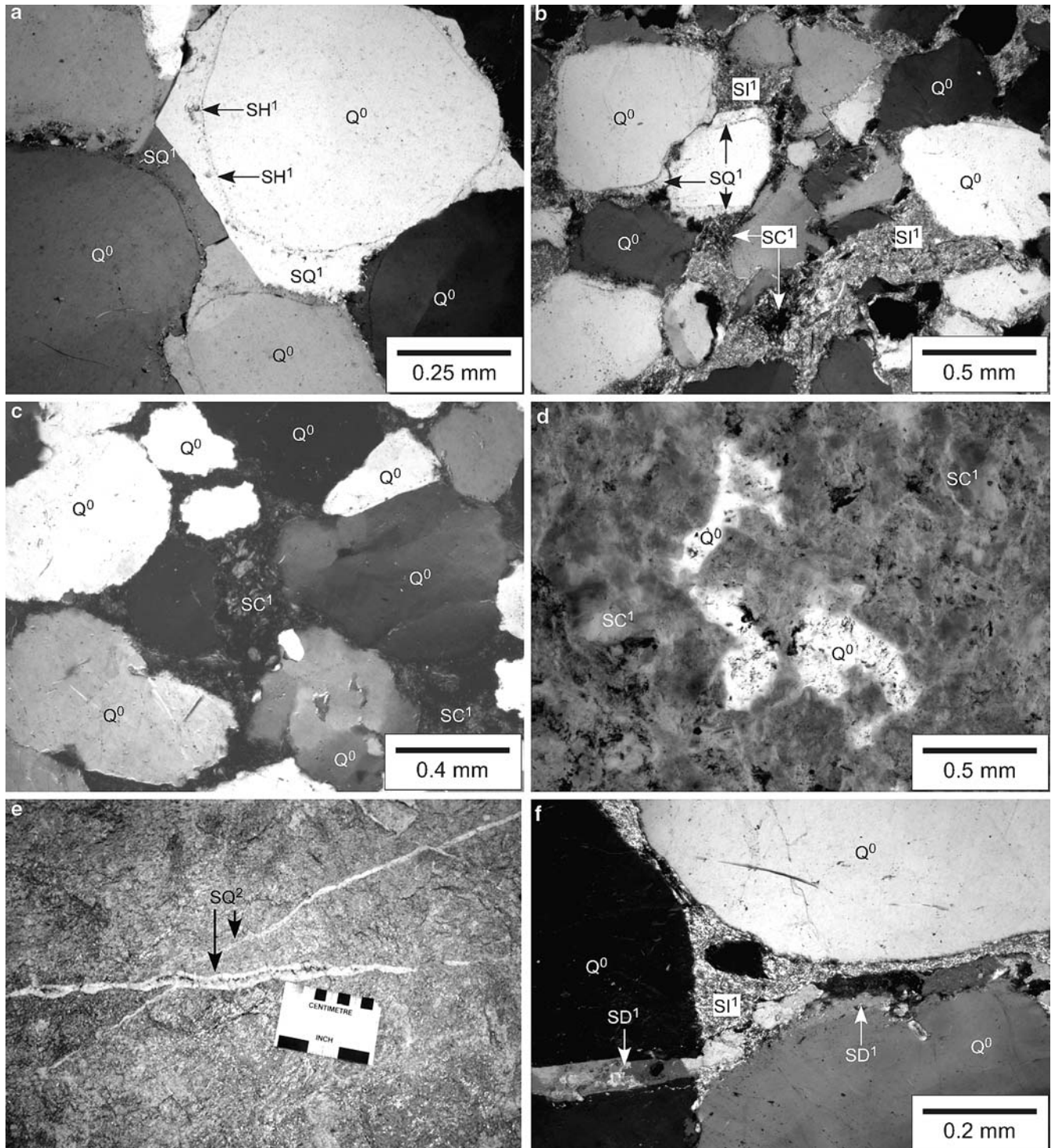
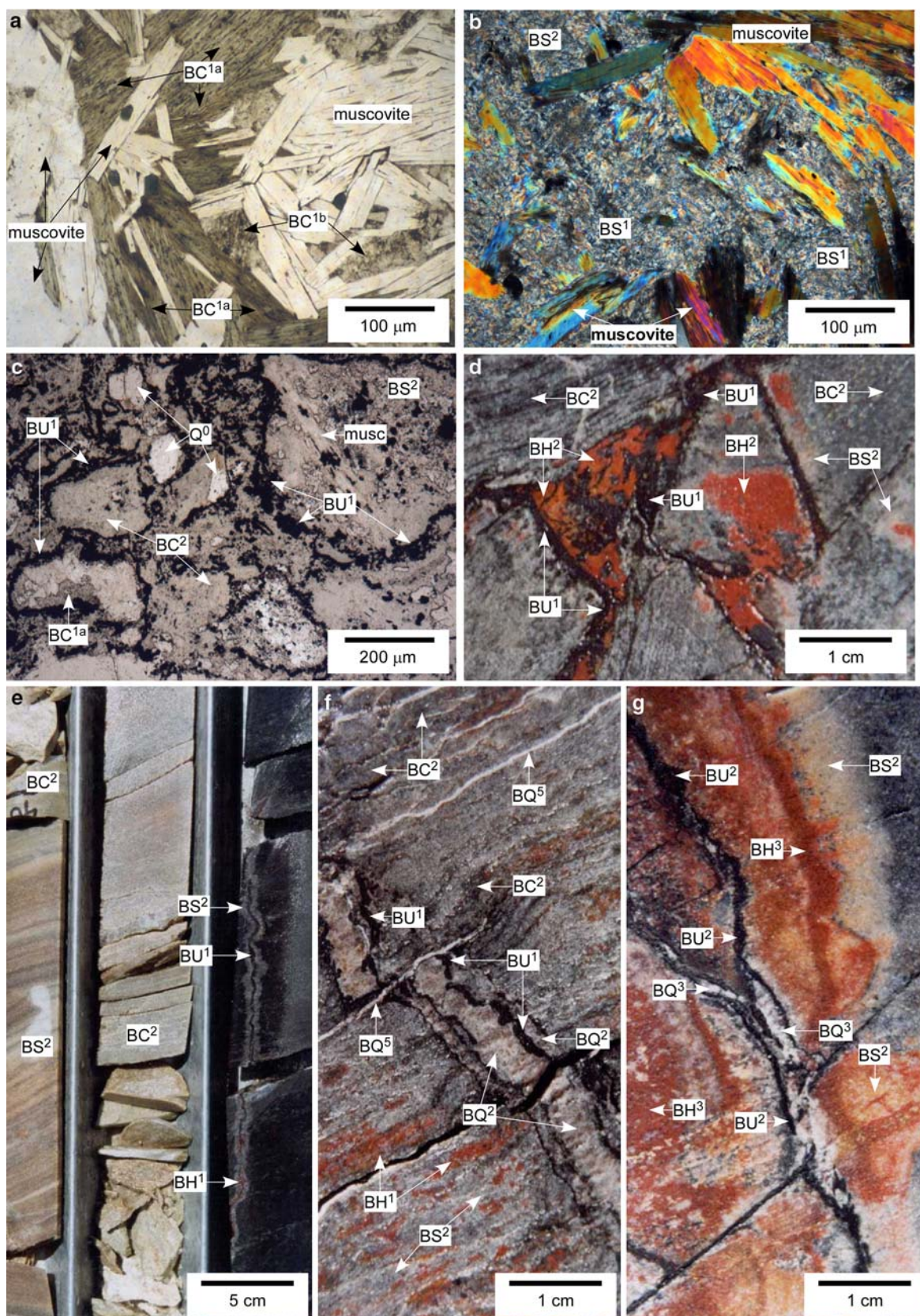


Fig. 8 a SH¹ hematite and trails of small, aqueous, fluid inclusions highlighting the boundary between detrital quartz (Q⁰) and SQ¹ quartz overgrowths. Sample N147V 151.1 m. b SI¹ replacing SQ¹ quartz overgrowths and SC¹ chlorite overprinting SI¹ illite in the Kombolgie Subgroup. Q⁰ = detrital quartz. Sample U189 W 137.2 m. c Pervasive SC¹ chlorite in the Kombolgie Subgroup. Q⁰ = detrital quartz. Sample S186 V 174.0 m. d Massive replacement of Q⁰ detrital quartz (and SI¹ illite) near the unconformity. Sample M141 N 210.9 m. e A 1 cm wide SQ² druzey quartz vein transecting lithified sandstones in the Kombolgie Subgroup on the surface above the Jabiluka II deposit. f A monomineralic SD¹ dolomite vein cutting SI¹ illite and detrital quartz grains. Sample S186 V 162.3 m

(Shepherd et al. 1985). Quartz veins, uraninite and phyllosilicate minerals from the deposit and overlying sandstone were extracted from a crushed and washed fraction of sample for stable isotope analysis. All samples were analysed by XRD using the method of Melling (1979) to ensure purity. In most cases, only pure separates of each phase were analysed. No sample contained > 5% contamination by another phase, usually chlorite or sericite. Oxygen isotopic compositions of

quartz, illite, sericite, chlorite, kaolinite and uraninite were measured using the BrF_5 method of Clayton and Mayeda (1963). Hydrogen isotopic compositions of

quartz-hosted fluid inclusions, sericite, chlorite, illite and kaolinite were determined using the methods of Kyser and O'Neil (1984). Stable isotopic measurements were



made using a Finigan MAT 252 mass spectrometer and are reported in the δ notation in units of per million relative to the standard V-SMOW. Replicate $\delta^{18}\text{O}$ analyses were reproducible to $\pm 0.2\text{‰}$ and δD values to $\pm 3\text{‰}$. Uraninite $\delta^{18}\text{O}$ values in two high-grade U samples were determined using the high spatial-resolution, CAMECA IMS 1270 ion microprobe at UCLA following the method of Fayek et al. (2000). Oxygen isotope fractionation factors used throughout this paper are those suggested by Sheppard and Gilg (1996) for water–illite, Land and Dutton (1978) for water–kaolinite, Yeh and Savin (1976) for water–chlorite, O’Neil and Taylor (1969) for water–muscovite, Clayton et al. (1972) for water–quartz and Fayek and Kyser (2000) for water–uraninite. Hydrogen isotope fractionation factors used are those suggested by Suzuoki and Epstein (1976) for water–muscovite, Yeh (1980) for water–illite, Lambert and Epstein (1980) for water–kaolinite and Graham et al. (1984, 1987) for water–chlorite.

Sericite, muscovite and illite separates were dated using the $^{40}\text{Ar}/^{39}\text{Ar}$ laser-heating technique of Lee et al. (1990). U–Pb and Pb–Pb analyses were performed using laser ablation (Mercantek LUV213 with Nd-YAG) coupled with an on-line Finnigan MAT Element high-resolution inductively coupled plasma mass spectrometer (LA-HR-ICPMS). The analyses were performed on polished thin sections in spot mode using the technique of Kyser et al. (2003). The isotopes measured were ^{201}Hg , ^{202}Hg (used for correction of interferences by ^{204}Hg on ^{204}Pb), ^{204}Pb (used for common Pb correction), ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{235}U , and ^{238}U .

Sandstone-hosted diagenesis and alteration

Sandstones above the Jabiluka deposit record a complex fluid history related to diagenesis, mineralization, and

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Fig. 9 **a** Plain transmitted light photomicrograph showing BC^{1a} chlorite pseudomorphing metamorphic biotite in the outer alteration zone, with apparently unaltered muscovite adjacent. Note also pore filling BC^{1b} chlorite. Sample N207V 546.7 m. **b** Polarised transmitted light photomicrograph showing fine-grained BS^1 sericite replacing metamorphic muscovite in the outer alteration halo. Sample N207 569.7 m. **c** Plain transmitted light photomicrograph showing rounded, chloritized schist fragments containing relict muscovite, biotite and quartz cemented by fine-grained, pale-brown and pale green BS^2 chlorite and disseminated BU^1 uraninite. Sample O186V 427 m. **d** Massive BU^1 uraninite and brick-red BH^2 hematite cementing angular carbonaceous schist clasts that are pervasively altered to BS^2 sericite and BC^2 chlorite. Sample JU37 60.2 m. **e** A BU^1 uraninite vein with a BS^2 sericite $\pm$ BH^1 hematite selvage in high angle fracture in carbonaceous schist. Sample JU31 44.5–46.0 m. **f** A BQ^2 quartz– BU^1 uraninite vein cross-cutting pelitic schists that have been altered by BC^2 chlorite, BS^2 sericite and BH^1 hematite. Note the preservation of the metamorphic texture in the schist and the displacement of the BQ^2 quartz– BU^1 uraninite vein by a BQ^5 quartz– BD^1 dolomite veinlet. Sample JU29 53.3 m. **g** Carbonaceous pelitic schist cross cut at a high angle by a BQ^3 quartz– BU^2 uraninite vein with a BH^2 hematite– BS^3 sericite selvage that obliterates the texture of the schist. Sample JU38 59.8 m

later tectonic events (Fig. 7). The event abbreviation nomenclature is described and illustrated in Fig. 7. The earliest preserved diagenetic event is represented by thin, discontinuous veneers of red–brown hematite (SH^1) that likely originated from altered Fe-hydroxides, and poor to well-developed syn-compaction quartz overgrowths ($\text{SQ}^{a/1}$) on detrital quartz grains (Fig. 8a). Compositionally mature (i.e. 100% quartz) sandstones that lacked detrital clay or feldspar, such as aeolian and upper shoreface facies, are marked by thick ($< 250\text{ }\mu\text{m}$), euhedral syntaxial quartz overgrowths on detrital grains that likely precipitated prior to significant mechanical and chemical compaction (SQ^a) (Hiatt et al. 2005). This cement was derived from paleoweathering or shallow ($< 1\text{ km}$) diagenetic processes and the alteration of interbedded volcanic rocks given that an extrinsic source of silica is necessary. A later stage of quartz cementation (SQ^1) associated with deeper burial, formed in open pores from quartz derived through compaction-driven, grain-to-grain pressure solution. SQ^a and SQ^1 quartz greatly reduced porosity, particularly in well-sorted, quartz-dominated lithologies, resulting in the formation of diagenetic aquitards.

Peak diagenesis began with the dissolution of detrital feldspar, smectitic clays, quartz overgrowths and detrital quartz grains in the proximal fluvial lithologies (Fig. 7). Lithologies that experienced such framework dissolution during deep burial were diagenetic aquifers in the Kombolgie Sub-basin (Hiatt and Kyser 2000; Kyser et al. 2000). The principle diagenetic mineral to precipitate during peak diagenesis is illite (SI^1), which commonly comprises 10–15% of the Kombolgie Subgroup in the diagenetic aquifer lithologies (Figs. 7, 8b). Tourmaline (ST^1) needles are rarely observed, but overprint SQ^1 quartz overgrowths and appear contemporaneous with SI^1 illite. SI^1 illite is overprinted in places by fine-grained pale-green to grey-green and olive-green chlorite (SC^1) (Fig. 8b), which occurs up to 312 m above the unconformity (Gustafson and Curtis 1983). Chlorite replacement of illite varies from limited amounts usually well above the unconformity, to pervasive in specific beds ten’s of metres above the unconformity (Fig. 8c) and ultimately as massive replacement of illite and detrital quartz, typically adjacent to the unconformity (Fig. 8d). Massive and pervasive chlorite intervals range between a few centimetres to several metres thick. Veins and veinlets of dark-green chlorite (SC^2) cross-cut SI^1 illite and SC^1 chlorite filled sandstone (Fig. 7). SC^2 chlorite-lined fractures and veins are common along the exposed unconformity of the decline (Fig. 5d) and are observed in many drill holes up 200 m above the unconformity. These SC^2 chlorite-lined fractures typically have a NNW to NW orientation and some may flatten and merge with bedding-controlled fractures. SC^2 chlorite appears to be contemporaneous with the SQ^2 druzy quartz and SH^2 hematite veins that also transect lithified sandstones and peak diagenetic minerals in the Kombolgie Subgroup in a NNW to NW orientation (Figs. 7, 5c, 8e). Pervasive non-destructive silica cement

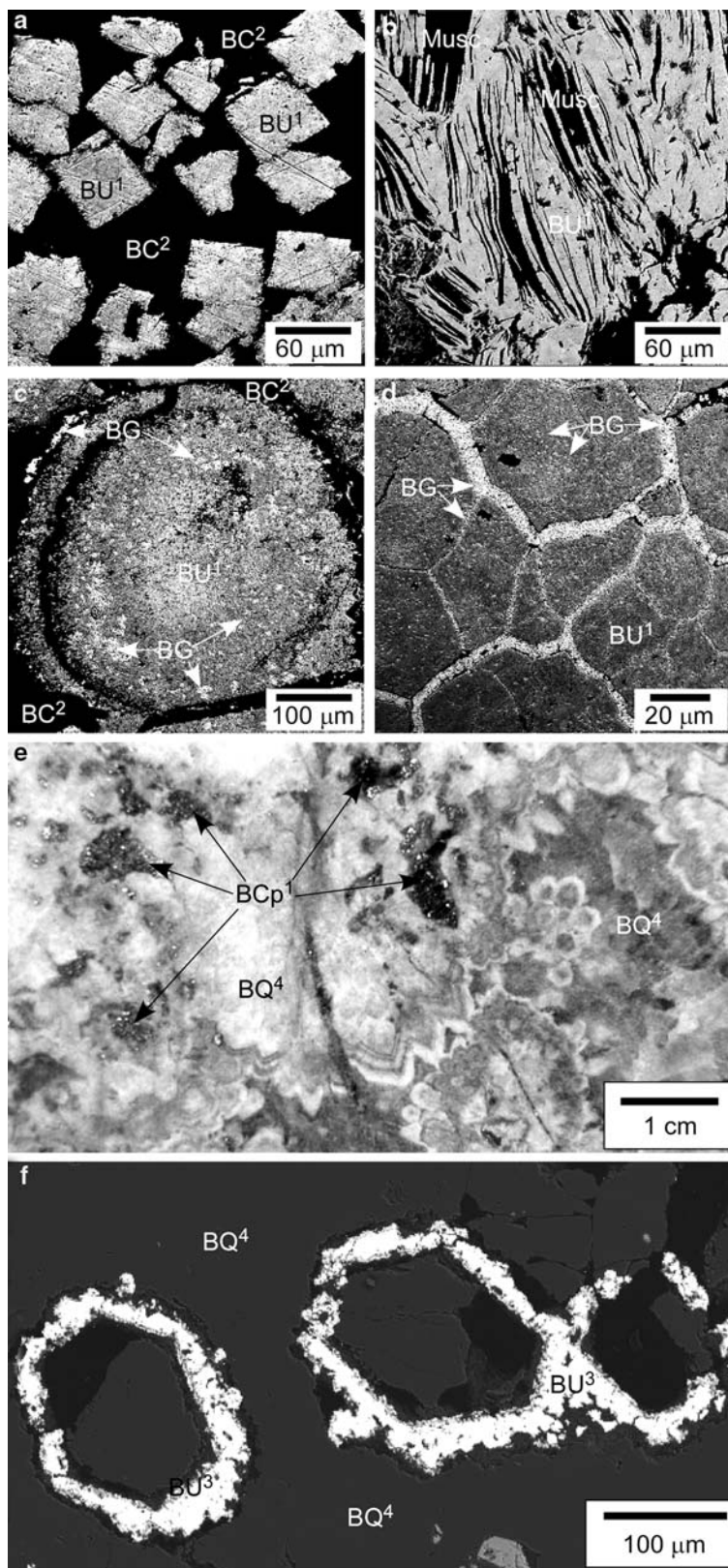
Fig. 10 **a** Backscatter image of cubic grains of disseminated BU¹ uraninite in a fine-grained BC² chlorite matrix.

b Backscatter image of BU¹ uraninite filling pores between metamorphic muscovite.

c Backscatter image of a small BU¹ uraninite botroid dotted with galena blebs (BG) and surrounded by BC² chlorite.

d A close-up backscatter image of a 5 cm wide massive BU¹ uraninite vein with no obvious inclusions in hand specimen containing shrinkage cracks and small voids filled with galena (BG). **e** A close-up photograph of a zoned, BQ⁴ drusy quartz vein with chalcopyrite blebs that cuts an unmineralized breccia zone. Sample JU31 12.5 m.

f Backscatter image of uraninite overgrowths on quartz in a BQ⁴ drusy quartz vein. Sample N147 V 398.7 m



optically similar to SQ¹ quartz, but correlated with SQ² quartz, locally overprints SC¹ chloritic sandstone, resulting in hard, chlorite-bearing rock. SQ³ drusy quartz ± SD¹ dolomite ± SP¹ pyrite ± SCp¹ chalco-

pyrite veins and monomineralic SD¹ dolomite veins are the last diagenetic minerals recorded in the sandstone and are likely a continuum of the SQ² quartz veins (Figs. 7, 8f).

Basement-hosted alteration

Jabiluka has an alteration halo in the basement rocks that extends at least 200 m beyond known uranium ore (Binns et al. 1980a, b), but a paucity of drill holes beyond these limits prohibits an accurate appraisal of its size. The earliest authigenic minerals in the basement lithologies preserved at Jabiluka are boudinaged and folded BQ¹ quartz ± muscovite veins (Fig. 5d) and thin, wispy foliation-parallel BP¹ pyrite stringers that occur in the graphitic to carbonaceous chlorite schists (Fig. 7). These minerals are now relicts of regional metamorphism and deformation. The BQ¹ quartz veins may have been conformable with S₁ foliation but were subsequently deformed during D₂ and D₃ forming tight to isoclinally folded 2–3 cm wide lenses and large ball-like boudins, up to 10 cm in diameter.

Alteration associated with uranium mineralization can be loosely divided into an outer and inner halo, with the principle difference being uranium mineralization and an increase in fine-grained chlorite and sericite associated with brecciation in the inner halo. This contrasts with the Nabarlek uranium deposit (Fig. 1) where the inner and outer alteration halos in the basement rocks are mineralogically distinct from one another (Wilde and Wall 1987; Polito et al. 2004). Alteration in the outer halo at Jabiluka is represented by chlorite (BC^{1a}) that pseudomorphs metamorphic biotite, garnet, amphibole, feldspar and sillimanite (Fig. 9a), fine-grained, pale-brown to pale-green clear chlorite (BC^{1b}) that occurs as void and matrix-filling material adjacent to the altered metamorphic minerals (Fig. 9a), fine-grained sericite (BS¹) that replaces metamorphic biotite, muscovite, amphibole and feldspar (Figs. 7, 9b) and traces of fine-grained red translucent BH¹ hematite. Metamorphic rutile needles that occur in relatively unaltered biotite laths transform to aggregates of BA¹ anatase during biotite alteration to chlorite (Fig. 7). Metamorphic zircon, muscovite, quartz and carbonaceous graphitic material are not visually altered in this area despite the alteration of adjacent phases (Binns et al. 1980a).

Within the mineralized areas, alteration becomes more intense and complex. Mineralized and unmineralized breccias comprising multiple chlorite and sericite generations ± quartz ± uraninite ± hematite exist. Mapping along with the 880 FW drift (Fig. 3) reveals that semi-brittle shears and breccias formed due to reverse movement. Semi-brittle shears are semi-conformable to the basement stratigraphy and appear to form a series of anastomosing slip planes that transect the stratigraphy and foliation at a low angle. Breccias are developed parallel to the gross stratigraphic layering (S₀) or conformable to the composite S₁/S₂ foliation. Breccias are also developed within the hinge zone of fault-related kink, chevron and box folds adjacent to reverse shears. Kink and chevron folds are developed best on the hanging wall side of reverse shears whereas box folds

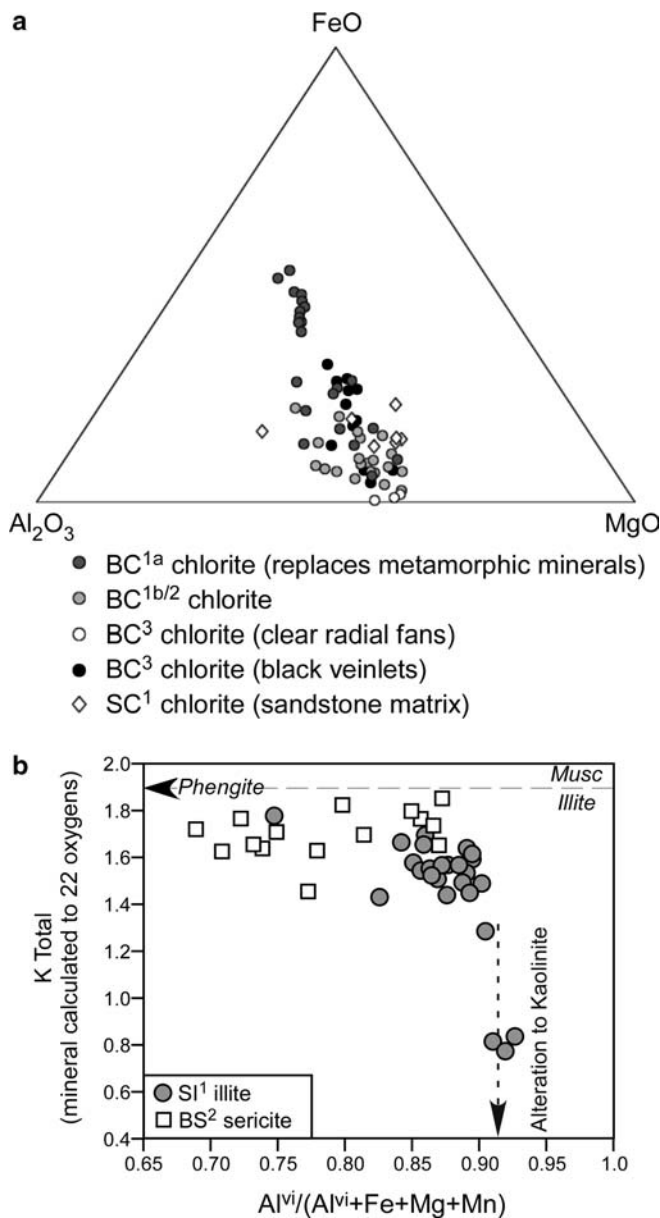


Fig. 11 **a** Ternary diagram showing the chemical variation of different chlorite types (in wt.%) around the Jabiluka deposit. **b** Relationship between total K and octahedral cation ratios in SI¹ illite from the Kombolgie Subgroup immediately overlying the Jabiluka deposit and BS² sericite from the lower Cahill Formation basement. Elements expressed in units per weight formula calculated to 22 oxygens. The muscovite (musc)–illite boundary is defined following the recommendation of Battaglia 2004). Alteration of SI¹ illites to kaolinite likely occurred during recent weathering

appear to be well developed on the footwall of structures. A similar pattern is also described from the Ranger 1 deposit (D. J. Johnston, unpublished PhD thesis). Both structures occur frequently in the carbonaceous laminated units and appear less common in the coarser grained, muscovite-rich units. The breccias appear to have good down-dip continuity up to 50 m, but display poor continuity along strike, with breccia inter-

vals mapped in the 880 Ore cross-cut unable to be correlated with those intercepted in drill sections centred at 48,925 and 48,940 m E, located 8 and 3 m either side of the drive respectively (Fig. 3).

Mineralized breccias typically comprise clast-supported frameworks of strongly chloritized and sericitized angular to rounded pelitic schist fragments cemented by fine-grained, dark-grey to pale-brown and green BC² chlorite and microscopic BU¹ uraninite (Fig. 9c) and brick-red BH² hematite $\pm$ BS² sericite $\pm$ BC² chlorite $\pm$ BQ² quartz (Figs. 7, 9d). BC² chlorite in the mineralized breccias is optically similar to BC^{1b} chlorite in the outer alteration, but whereas BC^{1b} chlorite passively fills voids between metamorphic minerals, BC² chlorite pervasively replaces most metamorphic minerals and obliterates textures. Nevertheless, it is likely that the two are related with BC² chlorite reflecting higher fluid/rock ratios around the breccias.

Cross-cutting relationships show that multiple generations of uraninite cut BC² chlorite and visa-versa. Uraninite also occurs with BS² sericite in high angle cross-cutting fracture-filled veins generally < 5 cm wide (Fig. 9e) and euhedral BQ² quartz in high angle cross-cutting veins (Fig. 9f). Graphite is obliterated in strongly bleached zones adjacent to massive uraninite veins, but is generally preserved particularly where BC² chlorite, BS² sericite and disseminated BU¹ uraninite comprise the mineralization. A later generation of monomineralic

uraninite veins (BU²) and BU² uraninite-quartz (BQ³) veins form an interconnected stockwork array that extends outward from some of the thicker subconcordant breccia intervals. These BU² uraninite $\pm$ BQ³ quartz veins occur with brick red finely-disseminated BH³ hematite and BS³ sericite that cross-cut and obliterate all earlier structures, metamorphic textures and minerals (Fig. 9g). The style and timing of this alteration is similar, but at a much smaller scale to that observed at Nabarlek where phengitic illite, red-purple hematite and uraninite obliterate all pre-ore textures and minerals, including quartz, in the mineralized zone (Polito et al. 2004).

A significant portion of the mineralization within the Main Mine Series occurs as microscopic grains that are easily overlooked, particularly in hand specimens rich in carbonaceous material. However, radiometric counts in excess of 1000 cps indicate the presence of finely disseminated uraninite. At the microscopic scale, uraninite appears as individual cubic or anhedral grains disseminated throughout a fine-grained phyllosilicate matrix (Fig. 10a), foliation-parallel pore filling cement between metamorphic muscovite (Fig. 10b) and small botroidal clusters frequently filled with galena blebs (BG) and BC³ chlorite (Fig. 10c). Petrographic studies also reveal that massive uraninite that has no obvious textures or inclusions in hand specimen contains shrinkage cracks and small voids filled with BG galena that may comprise

Table 2 Electron microprobe analyses of representative SI1 illite, BS2 sericite, BC1a/b chlorite, BC2 chlorite, BC3 chlorite and SC1 chlorite for Jabiluka

Sample	1	2	3	4	5	6	7	8
Oxide wt. %								
SiO ₂	32.3	25.1	33.3	32.4	31.6	30.0	48.6	52.7
Al ₂ O ₃	18.0	19.9	20.0	24.5	21.8	22.0	34.0	29.4
FeO	6.4	31.4	1.4	3.6	0.7	9.6	0.1	1.4
MnO	0.0	0.5	0.0	0.0	0.0	0.0	0.0	0.0
MgO	28.8	10.4	31.8	22.7	32.5	25.3	0.6	2.4
CaO	0.0	0.0	0.0	0.1	0.0	0.0	0.1	0.1
Na ₂ O	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.0
K ₂ O	0.0	0.0	0.2	0.1	0.0	0.1	9.1	10.0
Total	85.5	87.3	86.7	83.4	86.6	87.0	92.8	96.0
Calculated on the basis of 22 oxygens								
<i>Tetrahedral sites</i>								
Si	6.3	5.5	6.2	6.2	5.9	5.8	6.5	6.9
Al ^{IV}	1.7	2.5	1.8	1.8	2.1	2.2	1.5	1.1
Sum	8.0	8.0	8.0	8.0	8.0	8.0	8.0	8.0
<i>Octahedral sites</i>								
Al ^{VI}	2.4	2.7	2.6	3.8	2.6	2.8	3.8	3.4
Fe ²⁺	1.0	5.8	0.2	0.6	0.1	1.5	0.1	0.2
Mn	0.0	0.1	0.0	0.0	0.0	0.0	0.0	0.0
Mg	8.3	3.4	8.8	6.5	9.0	7.3	0.1	0.5
Sum	11.7	12.0	11.6	10.9	11.7	11.6	4.0	4.1
K ⁺	0.0	0.0	0.0	0.0	0.0	0.0	1.5	1.7
Ca ²⁺	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Na ⁺	0.0	0.0	0.0	0.0	0.0	0.0	0.1	0.0
Sum	0.0	0.0	0.0	0.0	0.0	0.0	1.6	1.7
Calc. Temp.	217	340	230	222	281	291	NA	NA

Sample 1 SC1 chlorite N207V 36.8 m; 2 BC1a chlorite R129V 149.4 m; 3 BC1b chlorite R129V 149.4 m; 4 BC2 chlorite R129V 199.0 m; 5 BC3 chlorite R129V 209.5 m; 6 BC3 chlorite R129V

117.5 m; 7 SI1 illite M141N 191.7 m; 8 BS2 sericite N207V 346.5 m; NA not applied; Calc. Temp. Calculated temperature of formation from chlorite chemistry

up to 10% of the sample (Fig. 10d). Brannerite (BB) forms < 5% of the ore and is likely coeval with uraninite (Fig. 7), but coffinite (BCo) crosscuts and replaces uraninite, indicating its later formation.

Unmineralized breccias crosscut mineralized breccias and are cemented by clear, zoned BQ⁴ drusy quartz and BC³ chlorite or a variable combination of very-fine grained BC³ chlorite ± zoned BQ⁴ quartz ± BS⁴ sericite ± BP² pyrite ± BCp¹ chalcopyrite ± BD¹ dolomite (Fig. 7). BC³ chlorite has numerous optical and textural properties that include transparent to pale-green radial fans that fill voids in post-ore breccias and dark-green and black veins that crosscut mineralized breccias. It is likely that these chlorite species did not form at the same time, but cross-cutting relations are lacking. Hand-held scintilometer readings of drill core reveal that un-

mineralized breccias generate < 500 cps compared to mineralized breccias that register > 1,000 cps. Zoned drusy quartz veins range from monomineralic comb-textured quartz (0.2–5.0 cm wide) to composite veins comprising zoned BQ⁴ quartz ± BP² pyrite ± BCp¹ chalcopyrite ± BD¹ dolomite (Fig. 10e). These veins have a N to NNE orientation and are texturally similar to that observed in the unmineralized breccias. The two are either related or represent a continuum from early-post to late-post-mineralization fluids. Tourmaline (BT) occurs as very fine-grained needle-like aggregates in some drusy quartz veins or as delicate concentric crustiform bands in the composite veins. Two composite BQ⁴ veins out of a total of 14 investigated were observed containing uraninite (BU³) as isolated angular fragments or as overgrowths around euhedral quartz (Fig. 10f).

Fig. 12 a The lower half of the sigmoidal BQ² quartz–BU¹ uraninite vein for fluid inclusion analysis. Note the growth zones highlighted by primary fluid inclusions in the quartz and the gradual change in uraninite morphology from massive, adjacent to the quartz vein, to disseminated, in the chlorite-dominated schist. Sample S132 V 204.1 m.

b Homogenization versus final melting plot for fluid inclusions in BQ⁴, BQ⁵ and SQ³ quartz veins from Jabiluka (Table 3). The grey shaded area represents the spread of values measured from fluid inclusions in quartz veins from throughout the Kombolgie Sub-basin by Derome et al. (2003a). Inset photographs show type A and B fluid inclusions in sample S132V 204.1 m. Note the elongate nature of the daughter mineral, which by its pale green colour is likely chlorite, not halite. DM daughter mineral, Vp vapour

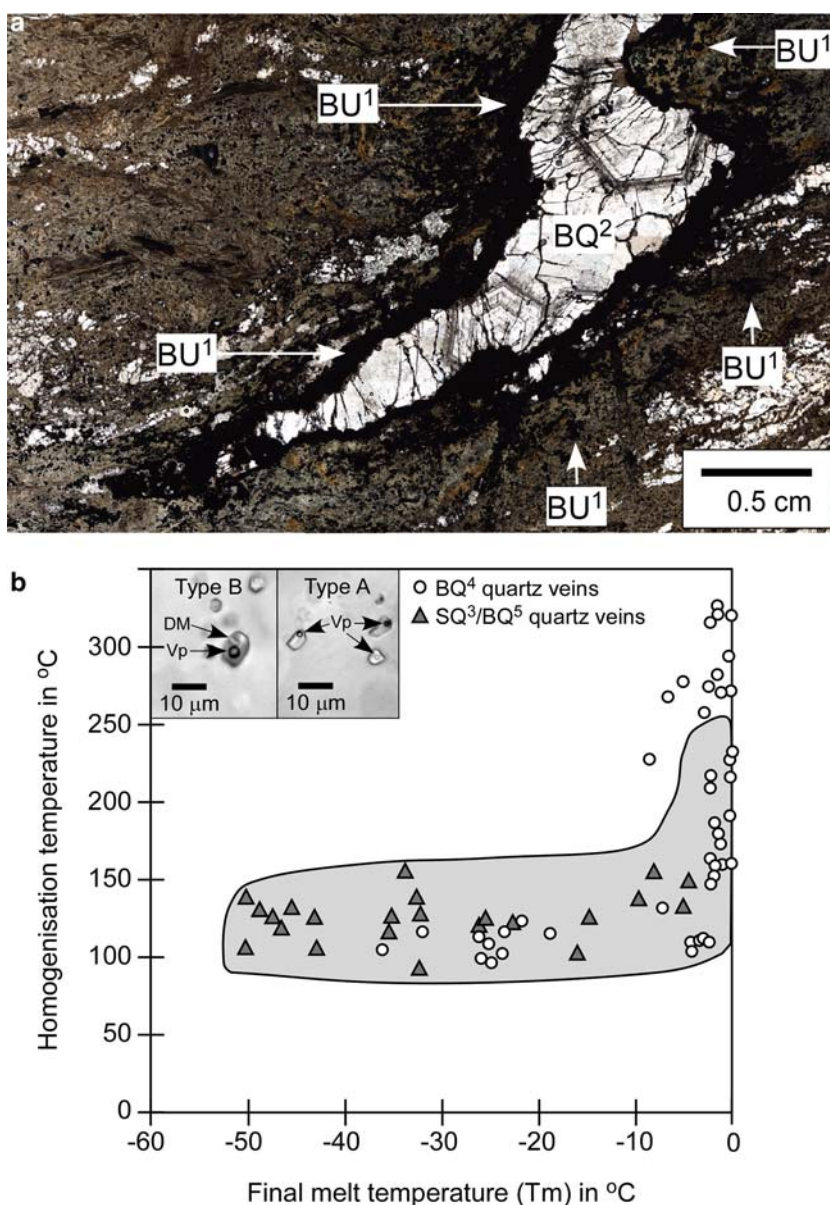


Table 3 Fluid inclusion microthermometry results from BQ2, BQ4, BQ5 and SQ3 quartz veins

Sample No.	Vein type	Vapour %	Phases	T_e	T_m	T_h
S132V 201.4 m	BQ ²	20	L and V	-37.3	-9.3	Decrepitated
S132V 201.4 m	BQ ²	10	L and V	-37.9	-11.6	Decrepitated
S132V 201.4 m	BQ ²	8	L and V	-38.9	-8.2	Decrepitated
S132V 201.4 m	BQ ²	12	L and V and S	-36.2	-17.1	Decrepitated
S132V 201.4 m	BQ ²	20	L and V	-30.5	-11.9	Decrepitated
S132V 201.4 m	BQ ²	15	L and V	-34.6	-12.2	Decrepitated
S132V 201.4 m	BQ ²	10	L and V and S	-39.0	-12.9	Decrepitated
S132V 201.4 m	BQ ²	10	L and V	-55.3	-13.2	Decrepitated
S132V 201.4 m	BQ ²	15	L and V	-40.1	-13.8	Decrepitated
S132V 201.4 m	BQ ²	10	L and V	-35.8	-12.0	Decrepitated
S132V 201.4 m	BQ ²	10	L and V	-29.9	-11.1	Decrepitated
M141N 244.1 m	BQ ⁴	12	L and V	-45.0	-8.8	227.5
M141N 244.1 m	BQ ⁴	10	L and V	-35.0	-0.1	231.5
M141N 244.1 m	BQ ⁴	8	L and V	-33.2	-6.8	267.4
N207V 524.3 m	BQ ⁴	15	L and V	-32.5	0.0	271.8
N207V 524.3 m	BQ ⁴	8	L and V	-25.5	-1.6	178.8
N207V 524.3 m	BQ ⁴	12	L and V	-28.6	-0.5	293.8
N207V 524.3 m	BQ ⁴	15	L and V	-30.1	-0.2	215.8
N207V 524.3 m	BQ ⁴	25	L and V	-32.0	-0.1	319.4
N207V 524.3 m	BQ ⁴	25	L and V	-30.1	-0.3	227.5
O186V 370.0 m	BQ ⁴	15	L and V	-26.1	-1.3	158.9
O186V 370.0 m	BQ ⁴	20	L and V	-38.0	-1.4	172.5
O186V 370.0 m	BQ ⁴	15	L and V	-32.5	-1.8	158.4
O186V 370.0 m	BQ ⁴	15	L and V	-30.1	-2.1	151.1
O186V 370.0 m	BQ ⁴	15	L and V	-28.6	-2.4	147.5
O186V 370.0 m	BQ ⁴	20	L and V	-29.8	-2.0	186.2
O186V 370.0 m	BQ ⁴	20	L and V	-24.6	-2.5	208.6
O186V 370.0m	BQ ⁴	25	L and V	-31.3	-2.3	216.7
O186 V 370.0m	BQ ⁴	15	L and V	-30.9	-1.3	270.0
O186 V 370.0m	BQ ⁴	20	L and V	-27.9	-2.4	314.9
O186 V 370.0m	BQ ⁴	30	L and V	-28.4	-1.8	325.1
O186 V 373.0m	BQ ⁴	5	L and V	-30.9	-2.4	163.2
O186 V 373.0m	BQ ⁴	5	L and V	-36.7	-2.6	109.7
O186 V 373.0m	BQ ⁴	8	L and V	-34.8	-3.1	112.2
O186 V 373.0m	BQ ⁴	8	L and V	-32.6	-3.7	110.5
O186 V 373.0m	BQ ⁴	5	L and V	-30.7	-4.0	108.4
O186 V 373.0m	BQ ⁴	5	L and V	-49.4	-4.3	103.3
O186 V 373.0m	BQ ⁴	5	L and V and S	-37.1	-2.0	Decrepitated
O186 V 373.0m	BQ ⁴	5	L and V and S	-28.7	-2.7	Decrepitated
O186 V 373.0m	BQ ⁴	5	L and V and S	-34.6	-2.8	Decrepitated
O186 V 373.0m	BQ ⁴	5	L and V and S	-38.2	-4.3	Decrepitated
O186 V 373.0m	BQ ⁴	20	L and V	-32.6	-3.1	256.8
O186 V 373.0m	BQ ⁴	20	L and V	-27.8	-2.5	273.7
O186 V 373.0m	BQ ⁴	50	L and V	-34.2	-5.2	277.5
O186 V 373.0m	BQ ⁴	10	L and V	-35.1	-1.6	281.2
O186 V 373.0m	BQ ⁴	20	L and V	-25.5	-1.6	320.4
O186 V 373.0m	BQ ⁴	25	L and V	-29.8	-2.0	Decrepitated
O186 V 373.0m	BQ ⁴	10	L and V	-30.1	-1.4	Decrepitated
O186 V 373.0m	BQ ⁴	8	L and V	-43.8	-23.8	102.1
O186 V 373.0m	BQ ⁴	5	L and V	-49.0	-23.7	116.6
O186 V 373.0m	BQ ⁴	5	L and V	-46.0	-25.0	96.4
O186 V 373.0m	BQ ⁴	10	L and V	-47.4	-25.3	109.0
O186 V 373.0 m	BQ ⁴	5	L and V	-42.1	-26.0	98.9
O186 V 373.0 m	BQ ⁴	8	L and V	-42.6	-23.7	Decrepitated
R129 V 122.4 m	BQ ⁴	10	L and V	-45.0	-7.3	131.6
R129 V 122.4 m	BQ ⁴	10	L and V	-47.7	-19.0	114.6
R129V 122.4 m	BQ ⁴	10	L and V	-50.9	-21.9	122.9
R129V 122.4 m	BQ ⁴	10	L and V	-69.4	-26.3	112.9
R129V 122.4 m	BQ ⁴	10	L and V	-64.2	-32.1	116.2
R129V 122.4 m	BQ ⁴	10	L and V	-68.0	-36.3	105.0
R129V 194.5 m	BQ ⁴	10	L and V	-32.0	-0.2	160.1
R129V 194.5 m	BQ ⁴	12	L and V	-30.1	-0.1	190.8
R129V 194.5m	BQ ⁴	15	L and V	-29.6	-1.3	Decrepitated
99J2-1	SQ ³	10	L and V	-67.5	-35.6	115.2
99J2-1	SQ ³	5	L and V	-56.8	-32.4	91.8
99J2-1	SQ ³	10	L and V	-64.8	-35.3	125.2
R189V 192.0m	SQ ³	10	L and V	-66.5	-22.8	121.0

Table 3 (Contd.)

R189V 192.0m	SQ ³	15	L and V	-59.9	-14.9	124.5
R189V 192.0m	SQ ³	5	L and V	-80.0	-43.2	124.7
R189V 192.0m	SQ ³	5	L and V	-67.5	-32.3	126.6
R189V 90.1m	SQ ³	5	L and V	-82.0	-46.7	117.4
R189 V 90.1m	SQ ³	10	L and V	-64.8	-26.2	119.8
R189 V 90.1m	SQ ³	10	L and V	-56.8	-25.6	124.1
R189 V 90.1m	SQ ³	10	L and V	-78.5	-47.6	124.7
R189 V 90.1m	SQ ³	10	L and V	-77.3	-45.6	130.4
R189 V 90.1m	SQ ³	10	L and V	-38.2	-4.6	148.1
R189 V 90.1m	SQ ³	10	L and V	-68.0	-32.7	137.4
R189 V 90.1m	SQ ³	10	L and V	-74.5	-48.9	129.4
R189V 90.1m	SQ ³	8	L and V	-47.4	-16.1	101.6
R189V 90.1m	SQ ³	5	L and V	-45.0	-5.2	131.9
R189V 90.1m	SQ ³	8	L and V	-49.4	-8.2	153.9
S186V 313.0m	BQ ⁵	15	L and V	-67.5	-33.9	154.6
S186V 313.0 m	BQ ⁵	12	L and V	-76.8	-43.0	104.7
S186V 313.0 m	BQ ⁵	10	L and V	-73.9	-50.3	104.8
S186V 313.0 m	BQ ⁵	10	L and V	-49.4	-9.8	136.7
S186V 313.0 m	BQ ⁵	10	L and V	-81.3	-50.3	138.4

T_e eutectic temperature in °C; T_m final melt temperature in °C; T_h homogenisation temperature in °C

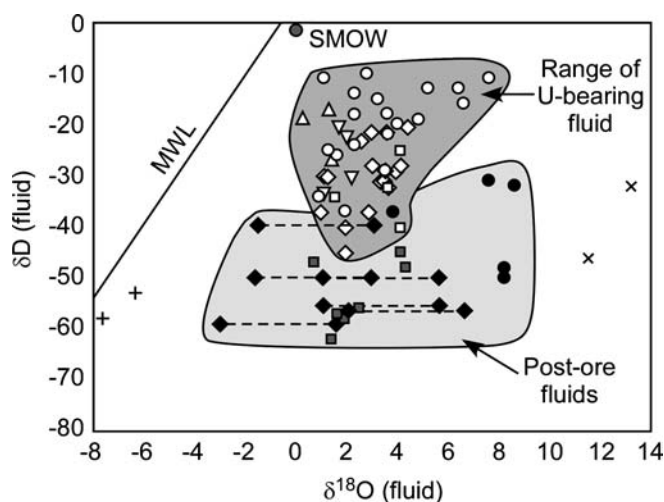
The isolated fragments were angular and it is interpreted that their occurrence is due to mechanical remobilisation of ruptured uraninite fragments. On the otherhand, uraninite that occurs as rims on euhedral quartz is interpreted as reprecipitation of local uraninite from the surrounding wall-rocks because no other post-ore veins contain this form of uraninite.

BQ⁵ drusy quartz $\pm$ BD² dolomite $\pm$ BP³ pyrite $\pm$ BCp² chalcopyrite veins up to 10 cm wide occur in the basement units and the overlying sandstone. In the sandstone, these veins are termed SQ³ quartz veins (Fig. 7). In the basement, these veins are straight sided with sharp wall-rock contacts and may extend for several metres along a development face. They crosscut all mineralized and unmineralized breccias and trend N to NNW, similar to the SQ³ quartz veins and SC² chlorite fractures. The quartz is always clear and formed prior to the dolomite and sulfides. The cross-cutting relationship with unmineralized breccias excludes these veins from being related to the BQ⁴ quartz in the breccias. Millimetre wide fibrous dolomite veins crosscut mineralized breccias and may have formed at the same time as the BQ⁵ quartz $\pm$ BD² dolomite $\pm$ sulfide veins.

Mineral chemistry

Seventy-seven chlorite analyses representing five different chlorite varieties were obtained by electron microprobe. SC¹ chlorite has a low Fe content and variable Al/Mg ratios (Fig. 11a). Temperature estimates calculated from this chlorite phase using the methods of Cathelineau and Nieva (1985) and Cathelineau (1988) indicate that this chlorite formed at $203 \pm 27^\circ\text{C}$ (Table 2). BC^{1b} and BC² chlorite has a composition similar to SC¹ chlorite (Table 2, Fig. 11a) and a calculated formation temperature of $224 \pm 26^\circ\text{C}$, indistinguishable

from SC¹ chlorite. BC^{1a} chlorite that pseudomorphs biotite, garnet, sillimanite and amphibole has a highly variable composition with respect to Fe and Mg (Fig. 11a). BC^{1a} chlorite that replaces biotite tends to be Fe-enriched, whereas chlorite replacing other metamorphic minerals tends to be Mg-enriched with a variable Al content (Fig. 11a). Temperature estimates calculated from C^{1b} chlorite are not reliable because they largely inherit the chemistry of the precursor mineral. Transparent, BC³ chlorite has a homogeneous composition



× Muscovite Δ BC^{1a} chlorite ◆ BQ⁴ quartz veins
 ◇ SI¹ illite ▽ BC² chlorite ■ BQ⁵/SQ³ quartz veins
 ○ SC¹ chlorite □ BS² sericite ● BC³ chlorite
 + Kaolinite

Fig. 13 Calculated $\delta^{18}\text{O}$ and δD values for various fluids associated with alteration at Jabiluka (Table 4). Tie lines represent the different $\delta^{18}\text{O}_{\text{fluid}}$ values calculated from the BQ⁴ quartz veins allowing for formation from mixed fluid with various temperatures. SMOW Standard mean ocean water, MWL meteoric water line

and low Fe content (Fig. 11a, Table 2) compared to the dark-green to black BC³ chlorite (Figs. 2, 11a), but both reveal formation temperatures of $284 \pm 6^\circ\text{C}$ and $280 \pm 36^\circ\text{C}$, respectively.

Electron microprobe analyses indicate that SI¹ illite is relatively pure, but some grains have FeO and MgO concentrations up to 3.4 and 1.3 wt.%, respectively reflecting a phengite composition (Table 2; Fig. 11b). Temperature estimates calculated from X-ray diffraction (XRD) patterns using the scheme developed by Kubler (1967) and Arkai (1991) indicate formation occurred at $200 \pm 50^\circ\text{C}$. BS² sericite has a slightly higher K total and lower $\text{Al}^{\text{IV}}/\text{Al}^{\text{IV}} + \text{Fe} + \text{Mg} + \text{Mn}$ concentrations than SI¹ illite and most compositions trend to that of phengite (Table 2; Fig. 11b). This is possibly because BS² sericite replaces muscovite and other Fe- and Mg-bearing metamorphic minerals from where it could inherit these elements. Previous researchers have shown that a higher K content corresponds to a higher formation temperature (Cathelineau 1988; Battaglia 2004) and seven crystallinity values from the BS² sericite by XRD analyses indicate that this sericite formed between 180 and 270°C , marginally higher than temperatures calculated from SI¹ illite. Five of these crystallinity values indicate formation at $200 \pm 20^\circ\text{C}$, which is in agreement with formation temperatures calculated from SI¹ illite and SC¹, BC^{1b} and BC² chlorite.

Fluid inclusions

Three previous fluid inclusion studies from Jabiluka (Ypma and Fuzikawa 1980; Wilde et al. 1989; Derome et al. 2003a) report that the uranium-bearing ore-fluids were highly saline, CaCl_2 – NaCl -dominated fluids. However, despite the presence of quartz–uraninite veins at Jabiluka (Fig. 9f, g), none of the previous studies analysed the fluid inclusions in these veins. Ypma and Fuzikawa (1980) report fluid inclusion analyses from a drusy quartz vein that is rimmed by and contains clasts of uraninite at Nabarlek, but the description of this vein fits that of our post-ore BQ⁴ quartz veins that contain isolated fragments of uraninite and have uraninite overgrowths (Fig. 10f). Wilde et al. (1989) analysed fluid inclusions in dolomite that is one of the last post-ore mineral phases to precipitate at Jabiluka (Fig. 8f) and Derome et al. (2003a) analysed fluid inclusions in metamorphic quartz veins that occur near mineralization and quartz-filled breccias and veins that are related to post-ore listric faulting at the unconformity along the decline to the 880 level.

Description of fluid inclusions

Fluid inclusions in one syn-ore BQ² quartz vein (Fig. 12a) and post-ore BQ⁴, BQ⁵ and SQ³ quartz veins were analysed during this study for the purpose

of comparing and contrasting the new data with previously published data. In this study, we describe two fluid inclusion types regardless of the vein analysed. Type A fluid inclusions contain <20 vol.% vapour and no daughter minerals (Fig. 12b). Type B fluid inclusions contain <20 vol.% vapour, halite and possibly one other daughter mineral or trapped solid (Fig. 12b). All of the fluid inclusions analysed in this study are found in randomly distributed three-dimensional clusters or along primary growth zones and are considered primary. Fluid salinities were not determined, because in most cases the final melting temperature was below the eutectic temperature of NaCl and it was rarely possible to distinguish CaCl_2 hydrate melting from hydrohalite or ice melting.

Microthermometry

Twelve fluid inclusions in one BU¹ uraninite–BQ² quartz vein were analysed by microthermometry. Two fluid inclusions contained one, pale-green trapped solid. The trapped solids were narrow and rectangular in one fluid inclusion and round in the other, characteristic of chlorite. Twelve fluid inclusions have a tight range of eutectic temperatures, with all but one inclusion melting at $-35 \pm 5.1^\circ\text{C}$ (Table 3) consistent with a mixture of CaCl_2 and NaCl . Final melting occurs between -17.1 and -8.2°C . Derome et al. (2003a) have demonstrated by using laser induced breakdown spectroscopy (LIBS) and Raman analysis that fluid inclusions with these microthermometric properties primarily occur in diagenetic quartz cements in the Kombolgie Subgroup and are dominated by Na^+ . Thus, our finding contrasts with claims that the syn-ore fluids involved a Ca-dominated component (Wilde et al. 1989; Derome et al. 2003a). None of the 12 fluid inclusions homogenised prior to decrepitation, making it impossible to calculate trapping temperatures. Such consistent decrepitation behaviour is difficult to explain in fluid inclusions that are not dominated by a gas phase and we propose that α -particle radiation damage may be responsible for weakening the crystal structure of the quartz since its formation. An alternative explanation is that N_2 , CH_4 , CO_2 , H_2 or O_2 may be present in the syn-ore fluid inclusions and could promote early decrepitation. These gases are recorded in post-ore drusy quartz veins in the ARUF (Derome et al. 2003a) and at the McArthur River uranium deposit in Canada (Derome et al. 2003b), but their presence does not generate widespread decrepitation in these studies when they are detected and it is unlikely that they are the cause for decrepitation here.

Six post-ore BQ⁴ drusy quartz veins were analysed. These veins contain type A and B fluid inclusions that record a wide range of chemical and thermometric properties. Four type B fluid inclusions were analysed, but decrepitation always occurred before homogenisation was reached. Type A fluid inclusions have eutectic temperatures that range from -69.4 to -24.6°C

Table 4 Oxygen and hydrogen stable isotope values of muscovite, quartz, chlorite, sericite, illite, uraninite and kaolinite varieties from the Jabiluka deposit

DDH	Paragenesis	$\delta^{18}\text{O}_{\text{min}}$	δD_{min}	Temp (°C)	$\delta^{18}\text{O}_{\text{fluid}}$	δD_{fluid}
JU37 66.4 m	Metamorphic muscovite	14.5	-50	500	13.2	-32
N207ND 335.9 m	Metamorphic muscovite	12.8	-64	500	11.5	-46
N147V 115.2 m	SI ¹ illite	11.3	-37	200	4.4	-21
N207ND 131.8 m	SI ¹ illite	8.9	-62	200	2.0	-46
N207ND 126.5 m	SI ¹ illite	8.9	-57	200	2.0	-41
N207ND 330.6 m	SI ¹ illite	9.8	-54	200	2.9	-38
N207ND 328.2 m	SI ¹ illite	10.2	-48	200	3.3	-32
R129V 72.3 m	SI ¹ illite	7.9	-54	200	1.0	-38
R129V 36.0 m	SI ¹ illite	8.0	-47	200	1.1	-31
R189V 114.9 m	SI ¹ illite	8.1	-47	200	1.2	-31
R189V 167.0 m	SI ¹ illite	9.5	-40	200	2.6	-24
S186V 158.2 m	SI ¹ illite	9.9	-45	200	3.0	-29
S186V 158.1 m	SI ¹ illite	10.5	-38	200	3.6	-22
V258ND 104.5 m	SI ¹ illite	10.5	-49	200	3.6	-33
X198V 71.6 m	SI ¹ illite	10.3	-48	200	3.4	-32
X198V 158.4 m	SI ¹ illite	10.4	-46	200	3.5	-30
X198V 121.9 m	SI ¹ illite	10.8	-46	200	3.9	-30
X198V 173.0 m	SI ¹ illite	11.0	-45	200	4.1	-29
X198V 86.5 m	SI ¹ illite	9.9	-38	200	3.0	-22
M141ND 208.8 m	SC ¹ chlorite	3.8	-55	200	1.6	-26
N147 V 59.8 m	SC ¹ chlorite	5.7	-58	200	3.5	-29
N1470 V 90.9 m	SC ¹ chlorite	4.5	-43	200	2.3	-14
N147 V 134.5 m	SC ¹ chlorite	3.1	-63	200	0.9	-34
N147 V 162.0 m	SC ¹ chlorite	4.1	-66	280	1.9	-37
N207 ND 43.9 m	SC ¹ chlorite	3.5	-54	200	1.3	-250
N207 ND 331.6 m	SC ¹ chlorite	6.2	-49	200	4.0	-20
N207 ND 17.8 m	SC ¹ chlorite	7.0	-48	200	4.8	-19
N207ND 193.8 m	SC ¹ chlorite	8.8	-45	200	6.6	-16
N207ND 331.6 m	SC ¹ chlorite	5.4	-44	200	3.2	-15
N207ND 36.8 m	SC ¹ chlorite	5.0	-39	200	2.8	-10
R129V 114.8 m	SC ¹ chlorite	4.5	-53	200	2.3	-24
R129V 86.7 m	SC ¹ chlorite	5.8	-51	200	3.6	-22
R129V 117.5 m	SC ¹ chlorite	3.3	-40	200	1.1	-11
R189V 187.4 m	SC ¹ chlorite	4.5	-47	200	2.3	-18
R189V 74.8 m	SC ¹ chlorite	5.8	-47	200	3.6	-18
V258ND 72.0 m	SC ¹ chlorite	7.4	-42	200	5.2	-13
V258ND 89.6 m	SC ¹ chlorite	9.8	-40	200	7.6	-11
V258ND 52.4 m	SC ¹ chlorite	8.6	-42	200	6.4	-13
M141ND 213.0 m	BC ¹ chlorite	2.5	-48	200	0.3	-19
M141ND 355.0 m	BC ¹ chlorite	3.6	-56	200	1.4	-270
N147V 410.7 m	BC ¹ chlorite	3.5	-46	200	1.3	-17
M141ND 213.0 m	BC ² chlorite	3.9	-49	200	1.7	-20
N147V 259.4 m	BC ² chlorite	4.2	-51	200	2.0	-22
N147V 391.2 m	BC ² chlorite	4.4	-59	200	2.2	-30
R129V 149.5 m	BC ² chlorite	3.3	-62	200	1.1	-33
N147V 288.5 m	BS ² sericite	10.2	-42	200	4.1	-25
N207ND 569.7 m	BS ² sericite	7.6	-51	200	1.5	-34
N207ND 334.8 m	BS ² sericite	9.7	-49	200	3.6	-32
Jabiluka 99J22	BS ² sericite	10.2	-57	200	4.1	-40
JU37 59.0 m	Massive Uraninite	-5.7	-	200	5.3	-
JU26 34.0 m	Massive Uraninite	-11.2	-	200	-0.2	-
JU37 59.0 m	Massive Uraninite	-9.3	-	200	1.7	-
JU37 58.8 m	Massive Uraninite	-11.7	-	200	-0.7	-
JU37 58.8 m	Massive Uraninite	-9.8	-	200	1.2	-
U135 V 179.3 m	Massive Uraninite	-9.5	-	200	1.5	-
N147 V 398.7 m	Massive Uraninite	-16.3	-	200	-5.3	-
N147 V 398.7 m	Vein Uraninite ^a	-11.9	-	200	-0.9	-
N147 V 398.7 m	Vein Uraninite ^a	-10.6	-	200	0.4	-
N147V 398.7 m	Massive Uraninite ^a	-17.6	-	200	-6.6	-
N147V 398.7 m	Massive Uraninite ^a	-18.7	-	200	-7.7	-
S186V 224.3 m	Massive Uraninite ^a	-16.0	-	200	-5.0	-
S186V 224.3 m	Massive Uraninite ^a	-13.0	-	200	-2.0	-
S186V 224.3 m	Massive Uraninite ^a	-13.0	-	200	-2.0	-
U189W 175.5 m	Botrioidal Uraninite ^a	-14.6	-	200	-3.6	-
U189W 175.5 m	Botrioidal Uraninite ^a	-21.7	-	200	-10.7	-
U189W 175.5 m	Botrioidal Uraninite ^a	-11.0	-	200	0.0	-

Table 4 (Contd.)

U189W 175.5 m	Botrioidal Uraninite ^a	−24.9	—	200	−13.9	—
U189W 175.5 m	Botrioidal Uraninite*	−21.1	—	200	−10.1	—
O186V 370.0 m	BQ ⁴ zoned drusy quartz vein	12.0	−59	150–225	−3.5 to 2.0	−59
N147V 261.8 m	BQ ⁴ zoned drusy quartz vein	13.5	−40	150–225	−2.0 to 3.5	−40
N207ND 388.0 m	BQ ⁴ zoned drusy quartz vein	16.3	−51	150–225	0.8 to 6.3	−51
N147V 353.6 m	BQ ⁴ zoned drusy quartz-pyrite vein	13.4	−50	150–225	−2.1 to 3.4	−50
N147V 304.4 m	BQ ⁴ zoned drusy quartz vein	16.1	−56	150–225	0.6 to 6.1	−56
Jabiluka 99J17	BQ ⁴ zoned drusy quartz vein	16.9	−57	150–225	1.4 to 6.9	−57
R189 V 192.0m	BQ ⁵ drusy quartz-dolomite vein	17.8	−45	170	4.1	−45
S186 V 313.0m	BQ ⁵ drusy quartz-dolomite-sulfide vein	15.6	−58	170	1.9	−58
N207 ND 304.6m	SQ ³ clear drusy quartz vein	15.1	−62	170	1.4	−62
N147 V 162.0m	SQ ³ clear drusy quartz vein	14.4	−47	170	0.7	−47
N147 V 146.8m	SQ ³ clear drusy quartz vein	15.3	−57	170	1.6	−57
N207 ND 61.0m	SQ ³ clear drusy quartz vein	16.2	−56	170	2.5	−56
X198 V 21.3m	SQ ³ clear drusy quartz vein	18.0	−48	170	4.3	−48
N147 V 258.4m	BC ³ chlorite	4.2	−65	280	3.8	−37
Jab Unmar	BC ³ chlorite	8.6	−76	280	8.2	−48
R129 V 209.4 m	BC ³ chlorite	8.6	−78	280	8.2	−50
N207ND 47.5 m	SC ² chlorite (in vein)	8.0	−59	280	7.6	−31
R189V 206.0 m	SC ² chlorite (at unconformity)	9.0	−60	280	8.6	−32
V258ND 15.5 m	Kaolinite	14.7	−75	40	−6.3	−53
X198V 8.0 m	Kaolinite	13.4	−80	40	−7.6	−58

Stable isotope values of the fluid from which these minerals precipitated are calculated using temperatures estimated in this paper (see text for details)

Abbreviations: $\delta^{18}\text{O}_{\min} = \delta^{18}\text{O}$ value of the mineral analyzed; $\delta D_{\min} = \delta D$ value of mineral analysed, *ND* not determined

^aUraninite = analyzed using a CAMECA IMS 1270 ion microprobe

(Table 3) and final melt temperatures range from −36.3 to 0.0°C (Fig. 12b). Homogenisation temperatures are equally varied and range from 103.3 to 325.1°C (Table 3). Derome et al. (2003a) report that drusy quartz veins that host fluid inclusions with identical chemical and thermal properties to those described here contain fluids dominated by Ca^{2+} and Na^+ with Ca:Na ratios ranging widely from 0.07 to 10.5. This range of chemical and thermal characteristics can be attributed to mixing between a Na-dominated and a Ca-dominated fluid as suggested by Derome et al. (2003a), but according to the paragenesis documented in the present study, this mixing occurred after uranium precipitation had ended and is not the cause for uraninite precipitation. A small population of fluid inclusions in these veins homogenise around $300 \pm 25^\circ\text{C}$ and complicate the mixing model. BC³ chlorite formed at the same time as the BQ⁴ drusy quartz veins (Fig. 7) and records formation temperatures of $280 \pm 36^\circ\text{C}$ (Table 2). Clearly these high temperatures are neither unique to a specific mineral phase nor can they be considered data outliers. These high temperatures suggest that a short-lived thermal spike, possibly coincident with the introduction of the Maningkorri/Mudginberri phonolitic dykes intruded near the deposit (Page et al. 1980) and was subsequently recorded.

Three post-ore SQ³ and one BQ⁵ quartz veins record one of the last, post-ore fluid events at Jabiluka. They trap fluid inclusions that have consistently low eutectic temperatures to −82.0°C and some very low final melting temperatures to −50.3°C (Table 3; Fig. 12b) consistent with a fluid dominated by Ca^{2+} (Derome et al. 2003a). These fluid inclusions cannot be distinguished

from the high salinity fluid inclusions trapped in the BQ⁴ quartz veins, suggesting that there may have been some overlap between the timing of these veins. Homogenisation occurs at $123 \pm 32^\circ\text{C}$, overlapping that observed in the BQ⁴ quartz veins.

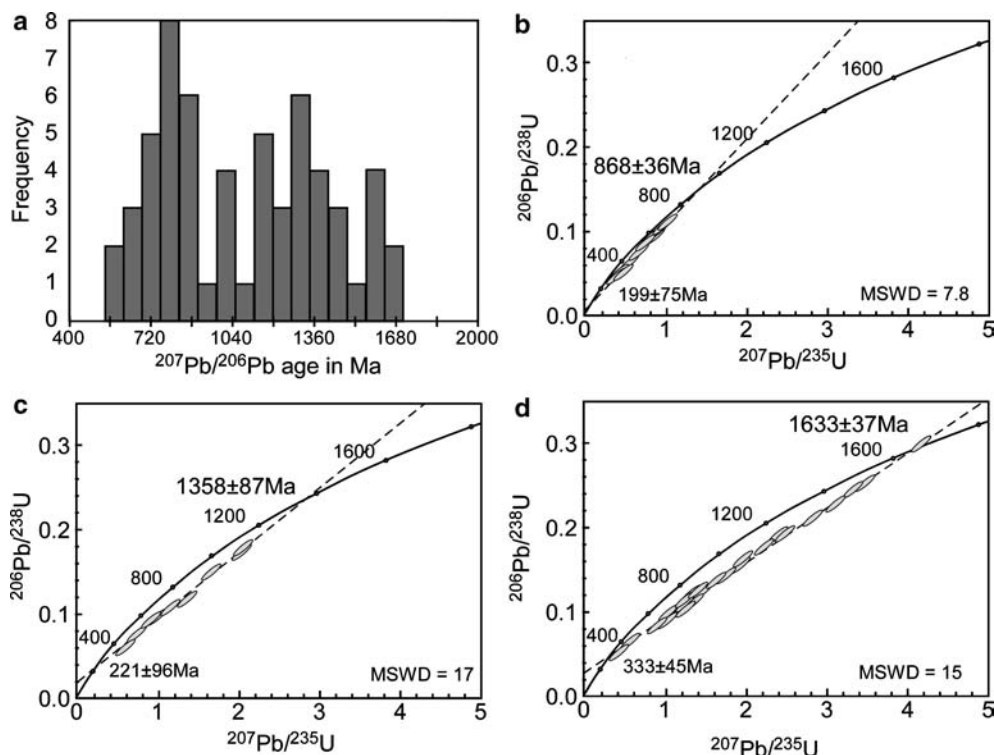
The microthermometric properties recorded in the post-ore quartz veins are indistinguishable from those reported by Wilde et al. (1989) and Derome et al. (2003a), whom both proposed that uranium precipitation was the result of the mixing between two contrasting fluids. The new results (Table 3) that are tied into a detailed paragenesis now suggest that this was not the case.

Stable isotopic compositions of formation fluids

Two metamorphic muscovite separates that represent the metamorphic mineral assemblage that reached middle to upper amphibolite grade during the Barra-mundi—Top End Orogeny, record $\delta^{18}\text{O}_{\text{fluid}}$ values of 11.5 and 13.2‰ and δD_{fluid} values between −46 and −32‰ (Table 4; Fig. 13). The high $\delta^{18}\text{O}_{\text{fluid}}$ values in particular are typical values for metamorphic fluids (Taylor 1997).

Oxygen isotope analysis by ion microprobe reveals that syn-taxial quartz overgrowths (SQ^a) in regional samples from the Kombolgie Sub-basin have $\delta^{18}\text{O}$ values that range from 16.1 to 21.1‰, whereas, the much more abundant, syn- to post-compaction quartz overgrowths (SQ¹) have lower $\delta^{18}\text{O}$ values ranging from 13.8 to 16.5‰ (Hiatt et al. 2005). The Kombolgie Sub-basin

Fig. 14 **a** Histogram of $^{207}\text{Pb}/^{206}\text{Pb}$ ages obtained from uraninite grains by LA-HR-ICPMS (Table 5). **b–d** U–Pb concordia diagrams from in situ isotopic analysis by LA-HR-ICPMS of uraninite from Jabiluka. Each plot represents a distinct population of uraninite formation, alteration or reprecipitation. The three discordia ages coincide with the majority of $^{207}\text{Pb}/^{206}\text{Pb}$ ages obtained. Plots were constructed using ISOPLOT (Ludwig 1993) from the isotope ratios presented in Table 5



formed at a low paleolatitude in a near-coastal setting (Idnurm and Giddings 1995; Sweet et al. 1999a, b), suggesting that these cements formed from meteoric water or seawater having a $\delta^{18}\text{O}_{\text{fluid}}$ of $0 \pm 4\text{‰}$. Using these $\delta^{18}\text{O}$ values, the depth at which the quartz cements formed is estimated at 1.9–4.2 km for SQ^a quartz and 4–5 km for SQ¹ quartz assuming a geothermal gradient of 30°C/km (Hiatt et al. 2005).

Sandstone-hosted SI¹ illite above Jabiluka has $\delta^{18}\text{O}_{\text{illite}}$ values from 7.9 to 11.3‰ and δD_{illite} from –62 to –37‰ (Table 4). The illite formed at ca. $200 \pm 50^\circ\text{C}$, indicating formation from a fluid with a $\delta^{18}\text{O}_{\text{fluid}}$ value between 1.0 and 4.4‰ and δD_{fluid} value of $-34 \pm 13\text{‰}$ (Table 4; Fig. 13). SC¹ chlorite has $\delta^{18}\text{O}_{\text{chlorite}}$ values from 3.3 to 9.8‰ and $\delta D_{\text{chlorite}}$ values from –58 to –33‰. SC¹ chlorite formed at ca. 200°C (Table 2) indicating formation from a fluid with a $\delta^{18}\text{O}_{\text{fluid}}$ value between 1.1 and 7.6‰ and a δD_{fluid} value of $-20 \pm 10\text{‰}$ (Table 4; Fig. 13). The $\delta^{18}\text{O}_{\text{fluid}}$ and δD_{fluid} values from the SI¹ illite and SC¹ chlorite have considerable overlap, plotting to the ^{18}O -rich side of the meteoric water line (Fig. 13), but having δD_{fluid} values near seawater. This implies that both phases precipitated from a basinal brine that evolved from a marine-dominated fluid with a minor meteoric component (Longstaffe 1987, 2000).

Pre- and syn-ore BC^{1a/b} and BC² chlorite has $\delta^{18}\text{O}$ values from 2.5 to 4.4‰ and δD values from –62 to –46‰ (Table 4). These values indicate formation in equilibrium with brines having $\delta^{18}\text{O}_{\text{fluid}}$ values of 0.3 to 2.2‰ and δD_{fluid} values of –33 to –17‰ based on a formation temperature of 200°C (Fig. 13). Four syn-ore BS² sericite samples with $\delta^{18}\text{O}_{\text{sericite}}$ values of 7.6 to

10.2‰ and $\delta D_{\text{sericite}}$ values between –57 and –42‰ (Table 4) formed in apparent equilibrium with a fluid having $\delta^{18}\text{O}_{\text{fluid}}$ values of 1.5 to 4.1‰ and δD_{fluid} values of –40 to –25‰ based on formation at ca. 200°C (Fig. 13), indistinguishable from the basement chlorites. Collectively, the pre- and syn-ore phyllosilicates from the Kombolgie Subgroup and the Lower Cahill Formation have a $\delta^{18}\text{O}_{\text{fluid}}$ value of $4.0 \pm 3.7\text{‰}$ and a δD_{fluid} value of $-27 \pm 17\text{‰}$.

Six $\delta^{18}\text{O}_{\text{uraninite}}$ values were obtained by micro-drilling pre-selected $< 2 \text{ mm}^2$ areas from thick sections of uraninite and analysing the powders by the fluorination technique described by Kotzer and Kyser (1991). The $\delta^{18}\text{O}_{\text{uraninite}}$ values range from –11.7 to –5.7‰ (Table 4). Thirteen other $\delta^{18}\text{O}_{\text{uraninite}}$ values were obtained during this study by high-resolution ion microprobe using a 20-μm beam and range between –24.9 and –10.6‰. Using the experimental uraninite–water fractionation factor for temperatures between 100 and 200°C (Fayek and Kyser 2000), the uranium is in equilibrium with fluids having a $\delta^{18}\text{O}_{\text{fluid}}$ value between –13.9 and +5.3‰ (Table 4). Only 5 of the 19 uraninite samples analysed have $\delta^{18}\text{O}_{\text{fluid}}$ values within the $4.0 \pm 3.7\text{‰}$ range calculated from the pre- to syn-ore phyllosilicate minerals, suggesting that only some uraninite has remained pristine since original precipitation, but most uraninite has re-equilibrated with fluids since then. The fourteen $\delta^{18}\text{O}_{\text{fluid}}$ values below 0.3‰ can only be explained by re-equilibration of the uraninite with meteoric fluids.

Post-ore BQ⁴ quartz veins have $\delta^{18}\text{O}_{\text{quartz}}$ values between 12.0 and 16.9‰. Fluid inclusions in the BQ⁴ veins

Table 5 LA-HR-ICPMS acquired U–Pb and isotopic data from uraninite samples from Jabiluka with calculated $^{235}\text{U}/^{207}\text{Pb}$, $^{238}\text{U}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages

Sample No.	Mineral	$^{207}\text{Pb}/^{235}\text{U}$	Isotopic ratios		$^{207}\text{Pb}/^{235}\text{U}$	Calculated ages (Ma) ^a	
			$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{206}\text{Pb}$		$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{206}\text{Pb}$
JU24 34.0 m	Uraninite	1.13382	0.11653	0.07196	770	711	985
JU24 34.0 m	Uraninite	0.4626	0.0536	0.0621	386	336	677
JU24 34.0 m	Uraninite	0.5499	0.0642	0.0624	445	401	687
JU24 34.0 m	Uraninite	3.3358	0.2432	0.0994	1489	1404	1612
JU24 34.0 m	Uraninite	3.1239	0.2271	0.1033	1439	1366	1685
JU37 36.2 m	Uraninite	0.76528	0.08818	0.06459	577	545	761
JU37 36.2 m	Uraninite	0.84813	0.09564	0.06630	624	589	816
JU37 36.2 m	Uraninite	0.88448	0.09705	0.06783	643	597	863
JU37 58.8 m	Uraninite	2.05018	0.17403	0.08575	1132	1034	1333
JU37 58.8 m	Uraninite	0.99054	0.09516	0.07671	699	586	1114
JU37 58.8 m	Uraninite	0.74994	0.07549	0.07225	568	469	993
JU37 58.8 m	Uraninite	0.60812	0.05938	0.07273	482	372	1006
JU37 58.8 m	Uraninite	1.16255	0.10645	0.07934	783	652	1181
JU37 58.8 m	Uraninite	0.92802	0.09236	0.07288	667	570	1011
JU37 58.8 m	Uraninite	1.27197	0.11267	0.08349	833	688	1281
JU37 58.8 m	Uraninite	1.46614	0.12426	0.08601	917	755	1338
JU37 58.8 m	Uraninite	0.92429	0.08234	0.07915	665	510	1176
JU37 58.8 m	Uraninite	1.35557	0.10969	0.08801	870	671	1383
JU37 58.8 m	Uraninite	1.26943	0.10377	0.08890	832	637	1402
JU37 58.8 m	Uraninite	1.64258	0.13655	0.09849	987	798	1596
JU37 58.8 m	Uraninite	1.04743	0.09174	0.08280	728	566	1265
JU37 58.8 m	Uraninite	1.06175	0.09789	0.07868	735	602	1164
JU37 58.8 m	Uraninite	1.33493	0.12260	0.07891	861	746	1170
JU37 58.8 m	Uraninite	1.24025	0.11344	0.07928	819	693	1179
JU37 58.8 m	Uraninite	2.49141	0.19172	0.09428	1270	1131	1514
JU37 58.8 m	Uraninite	1.94753	0.15531	0.09083	1098	931	1443
JU37 59.0 m	Uraninite	0.53835	0.06006	0.06424	437	376	750
JU37 59.0 m	Uraninite	2.05919	0.17581	0.08556	1135	1044	1328
JU37 59.0 m	Uraninite	1.36278	0.11642	0.08590	873	710	1336
JU37 59.0 m	Uraninite	1.66155	0.14884	0.08180	994	895	1241
JU37 59.0 m	Uraninite	2.83724	0.20929	0.09691	1365	1225	1566
JU37 59.0 m	Uraninite	4.18217	0.29849	0.10301	1671	1684	1679
JU37 59.0 m	Uraninite	2.23244	0.17754	0.09099	1191	1054	1446
JU37 59.0 m	Uraninite	2.40298	0.19207	0.09155	1244	1133	1458
JU37 59.0 m	Uraninite	3.48415	0.25269	0.10065	1524	1452	1636
JU37 59.0 m	Uraninite	1.42138	0.12583	0.08090	898	764	1219
JU37 59.0 m	Uraninite	1.77021	0.14296	0.08873	1035	861	1398
JU37 59.0 m	Uraninite	1.98472	0.16224	0.08754	1110	969	1373
JU37 59.0 m	Uraninite	1.25662	0.10858	0.08419	826	665	1297
JU38 41.3 m	Uraninite	1.16412	0.12373	0.06910	784	752	902
JU38 41.3 m	Uraninite	0.93235	0.10322	0.06620	669	633	813
JU38 41.3 m	Uraninite	0.99530	0.10782	0.06754	701	660	854
JU38 41.3 m	Uraninite	1.00864	0.11072	0.06767	708	677	859
R129V 207.9 m	Uraninite	0.94918	0.10598	0.06522	678	649	781
R129V 207.9 m	Uraninite	0.92112	0.10335	0.06491	663	634	771
R129V 207.9 m	Uraninite	0.99311	0.10876	0.06658	700	666	825
R129V 207.9 m	Uraninite	0.88767	0.09746	0.06631	645	600	816
R129V 207.9 m	Uraninite	1.01432	0.10798	0.06819	711	661	874
R129V 207.9 m	Uraninite	0.88250	0.09804	0.06545	642	603	789
R129V 207.9 m	Uraninite	0.87448	0.09119	0.06907	638	568	901
U135V 163.8 m	Uraninite	0.38968	0.04794	0.05944	334	302	583
U135V 163.8 m	Uraninite	0.45664	0.05444	0.06122	382	342	647
U135V 163.8 m	Uraninite	0.42291	0.05253	0.05858	358	330	552
U135V 179.0 m	Uraninite	0.68094	0.07706	0.06413	527	479	746
U135V 179.0 m	Uraninite	0.55512	0.06412	0.06286	448	401	704
U135V 179.0 m	Uraninite	1.02666	0.10756	0.06951	717	659	914
U135V 179.0 m	Uraninite	0.84123	0.09675	0.06320	620	595	715
U135V 179.0 m	Uraninite	1.01015	0.11179	0.06551	709	683	791

^a $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages calculated using equations reported by Ludwig (1993)

Table 6 Chemical composition and calculated chemical U-Pb ages from electron microprobe analyses of variably altered uraninite at Jabiluka

Sample	SiO ₂	PbO	UO ₂	FeO	CaO	TiO ₂	Total	Chem. age (Ma) ^a
JU37 58.8 m	0.3	5.6	87.8	0.2	0.9	0.1	95.1	397
JU37 58.8 m	0.3	5.9	88.5	0.2	0.8	0.1	95.9	416
JU37 58.8 m	0.3	6.0	88.5	0.2	1.0	0.1	96.1	417
JU37 58.8 m	0.3	7.3	87.3	0.3	0.9	0.1	96.1	518
JU37 58.8 m	0.3	5.6	89.0	0.3	1.0	0.1	96.3	393
JU37 58.8 m	0.4	5.5	88.8	0.2	1.0	0.1	96.1	384
JU37 58.8 m	0.5	5.2	84.2	0.3	1.1	0.1	91.5	385
JU37 59.0 m	0.3	6.1	87.2	0.2	1.0	0.1	95.0	436
JU37 59.0 m	0.3	6.3	87.4	0.2	0.8	0.2	95.1	448
JU37 59.0 m	0.4	6.0	90.4	0.2	1.1	0.1	98.3	412
JU37 59.0 m	0.3	6.0	85.0	0.2	1.0	0.1	92.6	435
N147 V 398.7 m	0.5	6.6	86.0	0.4	1.5	0.6	95.7	478
N147 V 398.7 m	0.6	6.7	85.3	0.4	1.6	0.6	95.1	490
N147 V 398.7 m	0.7	6.3	85.5	0.3	1.7	0.6	95.2	460
N147 V 398.7 m	0.5	6.8	86.9	0.5	1.4	0.4	96.5	487
N147V 398.7 m	0.7	6.3	86.2	0.4	1.8	0.4	95.8	451
N147V 398.7 m	0.7	6.1	87.5	0.3	1.7	0.3	96.6	434
N147V 398.7 m	0.7	6.0	87.6	0.4	1.6	0.3	96.7	427
O186V 224.3 m	0.4	5.9	88.4	0.3	1.0	0.1	96.1	416
O186V 224.3 m	0.2	6.1	88.4	0.2	1.0	0.2	96.1	427
O186V 224.3 m	0.5	5.7	88.3	0.1	1.2	0.1	95.9	400
O186V 224.3 m	0.3	6.4	88.6	0.2	0.9	0.2	96.6	451
O186V 224.3 m	0.4	5.9	88.5	0.2	1.3	0.1	96.3	415
O186V 224.3 m	0.3	6.3	88.2	0.2	1.2	0.2	96.3	442
O186V 224.3 m	0.3	6.5	87.8	0.3	1.3	0.2	96.4	457
O186V 224.3 m	0.2	6.0	88.0	0.2	1.1	0.2	95.8	421
O186V 244.5 m	0.4	10.0	82.7	0.3	1.7	0.4	95.6	747
O186V 244.5 m	0.5	10.1	81.5	0.4	1.7	0.6	94.8	767
O186V 244.5 m	0.4	9.7	82.2	0.5	1.7	0.5	94.9	730
O186V 244.5 m	0.4	10.5	81.6	0.5	1.6	0.4	94.9	796
O186V 244.5 m	0.4	11.9	81.7	0.4	1.6	0.5	96.6	903
O186V 244.5 m	0.4	10.4	82.5	0.3	1.6	0.5	95.7	786
O186V 244.5 m	0.3	11.5	81.8	0.3	1.3	0.6	95.8	871
O186 V 244.5 m	0.4	9.9	82.5	0.4	1.3	0.8	95.2	744
O186 V 244.5 m	0.4	12.2	80.6	0.3	1.6	0.5	95.6	940
O186 V 244.5 m	0.4	11.5	81.3	0.4	1.5	0.5	95.8	880
O186 V 244.5 m	0.4	11.9	80.7	0.4	1.5	0.5	95.6	912
O186 V 244.5 m	0.5	9.0	81.9	0.4	1.9	0.5	94.1	679
O186 V 244.5 m	0.4	11.2	80.7	0.4	1.8	0.7	95.2	864
O186 V 244.5 m	0.4	7.8	82.4	0.4	2.2	0.6	93.8	589
O186V 244.5 m	0.5	9.1	80.5	0.5	2.0	0.8	93.4	698
O186V 244.5 m	0.5	9.8	81.5	0.4	1.9	0.7	94.8	745
R129V 207.9 m	0.4	7.0	85.9	0.3	1.7	0.6	95.9	503
R129V 207.9 m	0.4	7.7	85.5	0.3	1.5	0.5	96.0	562
R129V 207.9 m	0.4	7.6	85.7	0.3	1.6	0.6	96.2	552
R129V 207.9 m	0.3	7.6	84.8	0.2	1.6	0.5	95.2	556
R129V 207.9 m	0.3	7.7	85.9	0.3	1.7	0.5	96.4	555
R129V 207.9 m	0.3	8.3	84.6	0.3	1.4	0.5	95.4	608
R129V 207.9 m	0.4	8.8	85.4	0.3	1.4	0.5	96.7	641
R129V 207.9 m	0.4	8.1	85.5	0.2	1.5	0.6	96.3	591
U135V 163.8 m	0.6	6.7	85.9	0.4	1.6	0.6	95.8	487
U135V 163.8 m	0.7	6.4	86.3	0.4	1.7	0.6	96.1	462
U135V 163.8 m	0.6	7.0	86.1	0.3	1.5	0.6	96.1	504
U135V 163.8 m	0.7	7.5	86.2	0.3	1.4	0.5	96.6	537
U135V 163.8 m	0.8	6.3	86.2	0.4	1.9	0.6	96.2	456
U135V 163.8m	0.7	6.0	86.9	0.3	1.8	0.2	96.0	428
U135V 163.8 m	0.9	5.7	87.0	0.3	1.9	0.3	96.1	410
U135V 163.8 m	0.7	6.0	87.8	0.3	1.6	0.2	96.7	427
U135V 179.3 m	0.4	6.2	85.4	0.4	1.7	0.0	94.2	452
U135V 179.3 m	0.4	6.3	85.7	0.3	1.3	0.0	94.0	459
U135V 179.3 m	0.3	6.3	87.5	0.3	1.1	0.1	95.5	449
U135V 179.3 m	0.8	5.6	86.1	0.3	1.9	0.0	94.6	402
U135V 179.3 m	0.2	6.7	87.1	0.3	1.1	0.0	95.3	474
U135V 179.3 m	0.7	5.7	86.8	0.3	1.5	0.1	95.1	409
U135V 179.3 m	0.4	6.3	86.4	0.2	1.2	0.0	94.5	455
U135V 179.3 m	0.3	6.8	86.2	0.2	1.0	0.1	94.5	488

Table 6 (Contd.)

Sample	SiO ₂	PbO	UO ₂	FeO	CaO	TiO ₂	Total	Chem. age (Ma) ^a
U135V 179.3 m	0.4	6.2	87.0	0.3	1.3	0.0	95.2	445
U135 V 179.3m	0.4	5.9	87.4	0.2	1.3	0.0	95.3	421

^aChemical age calculated using the equation: Age in Ma = $Pb \times 10^4 / (1.612 U)$ (Bowles 1990).

record a wide range of homogenisation temperatures that reflect entrapment of two distinct fluids. Derome et al. (2003a) calculated a minimum and maximum formation temperature of 150°C and 225°C for these veins, excluding higher temperature fluid inclusions as being an anomalous population. At the temperatures calculated by Derome et al. (2003a) the BQ⁴ veins precipitated from fluids having $\delta^{18}O_{\text{fluid}}$ values that may have been as low as -3.4 and as high as $+6.9$ ‰. Regardless of temperature, the same veins formed from fluids having δD_{fluid} values between -59 and -40 ‰ (Table 4; Fig. 13). Post-ore BC³ and SC² chlorite has $\delta^{18}O_{\text{chlorite}}$ values from 4.2 to 9.0 ‰ and $\delta D_{\text{chlorite}}$ values

from -78 to -59 ‰ indicating formation in equilibrium with fluids having $\delta^{18}O_{\text{fluid}}$ values of 3.4 – 8.2 ‰ and δD_{fluid} values of -50 to -31 ‰ based on a formation temperature of 280°C (Table 2). BQ⁵ and co-existing SQ³ quartz $\pm$ dolomite veins have $\delta^{18}O_{\text{quartz}}$ values between 14.4 and 18.0 ‰ and formed at 170 ± 30 °C. These veins formed from brines with $\delta^{18}O_{\text{fluid}}$ values between 0.7 and 4.3 ‰ and δD_{fluid} values between -62 and -45 ‰ (Table 4; Fig. 13). These results show that the post-ore minerals have isotopic values that are distinct from those recorded in the syn-ore minerals (Fig. 13). The post-ore range of $\delta^{18}O_{\text{fluid}}$ and δD_{fluid} values indicates formation from a variably evolved meteoric fluid with a relatively minor contribution from a marine source (Longstaffe 1987, 2000).

Two kaolinite separates extracted from the Kombolgie Subgroup have $\delta^{18}O_{\text{kaolinite}}$ values between 13.4 and 14.7 ‰, and $\delta D_{\text{kaolinite}}$ values between -80 and -75 ‰. If this kaolinite formed at 30 ± 10 °C in accordance with equatorial temperatures during weathering, it formed in equilibrium with fluids having a $\delta^{18}O_{\text{fluid}}$ value of -7 ± 0.6 ‰ and a δD_{fluid} value of -55 ± 3 ‰ (Fig. 13), equivalent to values obtained from modern meteoric water in northern Australia.

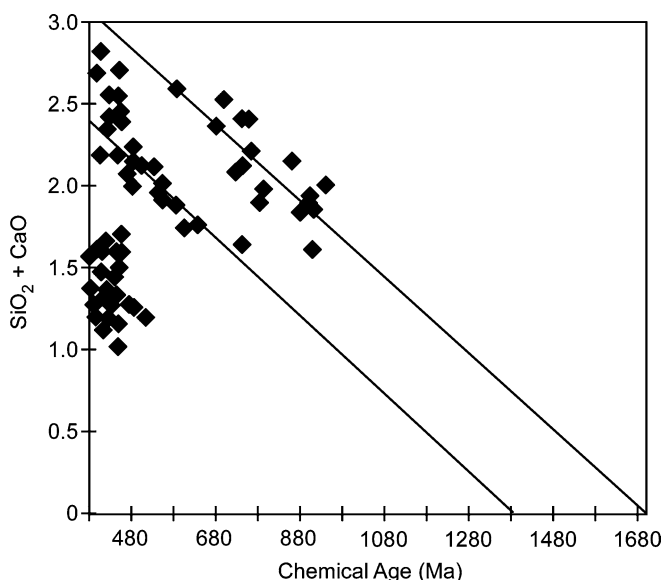
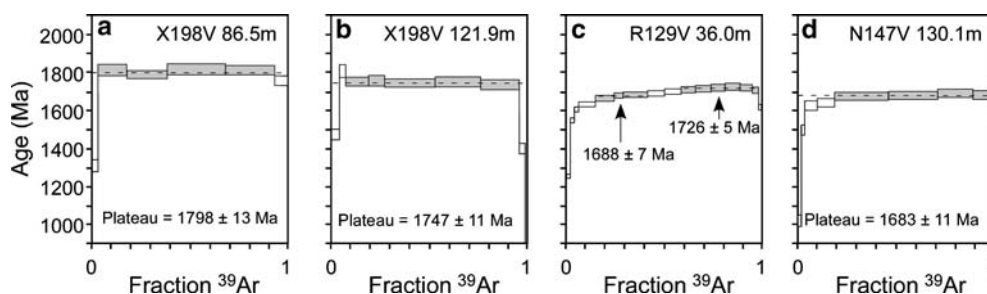


Fig. 15 Relationship between U/Pb chemical ages and SiO₂ + CaO contents in uraninite (Table 5) showing three populations defining at least two ages that correspond to $^{207}\text{Pb}/^{206}\text{Pb}$ ages and major regional tectonic events

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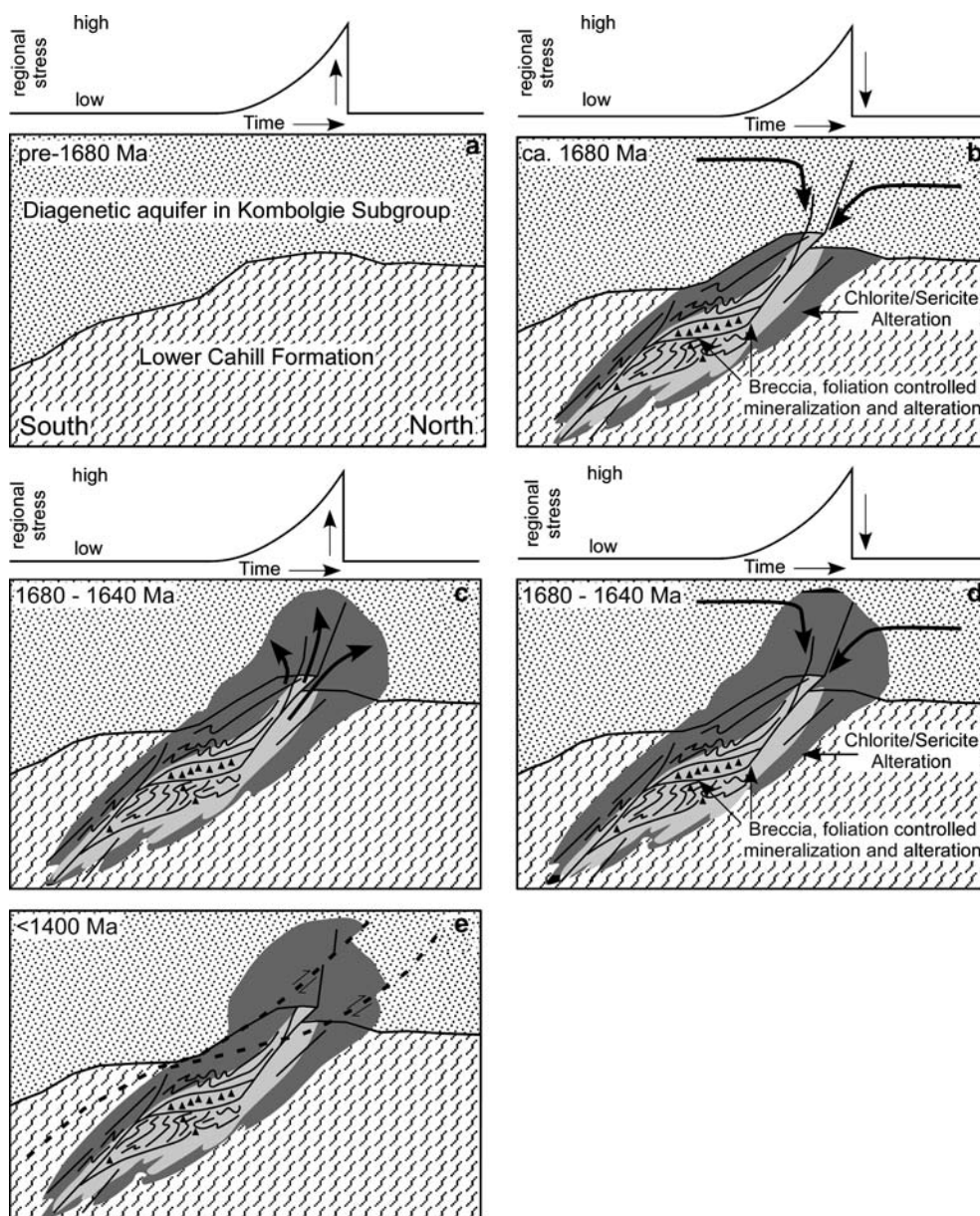
Uraninite grains analysed by LA-HR-ICPMS have $^{207}\text{Pb}/^{206}\text{Pb}$ ages that range from 552 to 1,685 Ma (Table 5). All of the samples analysed have a mottled appearance with micron-size areas of highly reflective uraninite typically surrounded by less reflective and dull uraninite (Fig. 10c, d). Galena that likely originated by the radioactive decay of uranium commonly comprises 5–10% of some samples and implies Pb loss and diffusion out from the uraninite. A significant proportion of the $^{207}\text{Pb}/^{206}\text{Pb}$ ages occur at $ca. 802 \pm 87$ Ma, with another major peak occurring around $1,302 \pm 37$ Ma

Fig. 16 $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra for four representative samples of SI¹ illite from the Kombolgie Formation at Jabiluka. **a** X198V 86.5 m. **b** X198V 121.9 m. **c** R129V 36.0 m. **d** N147V 130.1 m. Complete step heat data and interpreted ages are provided in Table 7



(Fig. 14). The oldest $^{207}\text{Pb}/^{206}\text{Pb}$ ages of $1,679 \pm 17$ and $1,685 \pm 17$ Ma are amongst the oldest ages from uranium mineralization in the ARUF. These ages come from areas within a uraninite grain having the highest reflectivity. Previous investigations have demonstrated that highly reflective uraninite has typically lost the least amount of Pb and gives the oldest ages (Kotzer and Kyser 1993; Fayek and Kyser 1997; Fayek et al. 2000; Polito et al. 2004). One of these ages corresponds to an area in sample JU37 59.0 m that has a $\delta^{18}\text{O}_{\text{fluid}}$ value of 5.3‰, indicating that it has experienced minimal post-precipitation alteration. The apparent episodic resetting of uraninite recorded by $^{207}\text{Pb}/^{206}\text{Pb}$ ages younger than 1,679 Ma, occurs in less reflective uraninite and indicates that the U–Pb isotope system has been partially to completely disturbed by multiple, fluid-related events since original formation.

Three U–Pb discordia diagrams reveal upper intercept ages that are similar to the $^{207}\text{Pb}/^{206}\text{Pb}$ ages at 888 ± 52 , $1,358 \pm 87$ and $1,633 \pm 37$ Ma (Fig. 14b–d). Lower intercept ages for each isochron are around 268 ± 144 Ma, with the high-error margin attributed to data scatter in this range. The previously published U–Pb discordia age of $1,437 \pm 18$ Ma from Jabiluka (Ludwig et al. 1987) assumed an isotopically closed system, but our results indicate that an age of 1,437 Ma reflects mixing between uraninite that preserves its original formation age and uraninite that was reset by younger fluid events. The majority of our U–Pb isotope ratios are significantly discordant, a phenomenon that is well documented in unconformity-related uranium deposits in Australia and Canada (Hills and Richards 1976; Ludwig et al. 1987; Kotzer and Kyser 1993; Fayek and Kyser 1997; Fayek et al. 2000; Polito et al. 2004). Unlike



most zircon U–Pb data, uraninite U–Pb data is characterized by discordance, high error margins, high MSWD values (Fig. 14) and mottling (Fig. 10c, d). This is primarily due to the refractive nature of the uraninite (Romberger 1984) and the variable amounts of Pb that can be lost from the uraninite after formation. Errors can be compounded by the LA-HR-ICPMS spot size of 25 µm being larger than the mottling within heterogeneous uraninite and in cases where ablation pits overlap two stages of uraninite within a single thin section intermediate $^{207}\text{Pb}/^{206}\text{Pb}$ ages can be expected. Indeed, mottled uraninite samples can produce $^{207}\text{Pb}/^{206}\text{Pb}$ ages that differ by as much as 1,000 million years (Table 5) and for these reasons, it is unlikely that a single uraninite sample will necessarily preserve U–Pb ratios that form an isochron to define only one age. The oldest $^{207}\text{Pb}/^{206}\text{Pb}$ ages are interpreted to be the closest to the original formation age for the Jabiluka deposit because $^{207}\text{Pb}/^{206}\text{Pb}$ ages in uraninite are more robust than U–Pb ages (Kotzer and Kyser 1993; Kyser et al. 2003). This is because the former uses isotope ratios of the same element, whereas U and Pb concentrations and therefore their respective ratios can be vastly different in altered uraninite (Burns 1999).

Electron microprobe data (Table 6) and backscattered electron images (Fig. 10) from seven uraninite-bearing samples show that Jabiluka uraninite has a variable composition, comprising 80.5–90.4 wt.% UO_2 , 5.2–12.2 wt.% PbO, 0.8–2.2 wt.% CaO, 0.1–0.5 wt.% FeO and 0.2–0.9 wt.% SiO_2 . The U:Pb ratios determined from their respective oxide proportions correlate

to chemical ages that range from 384 to 940 Ma (Fig. 15) using Eq. 1 (Bowles 1990).

$$\text{Age in Ma} = \frac{\text{Pb} \times 10^4}{1.612\text{U}}. \quad (1)$$

Naturally occurring, newly formed uraninite has a chemical composition that ranges from $\text{UO}_{2.07}$ to $\text{UO}_{2.25}$ (Burns 1999) and < 0.1 wt.% CaO, FeO and SiO_2 (Bowles 1990). The oldest $^{207}\text{Pb}/^{206}\text{Pb}$ ages (Table 6) indicate that Jabiluka uraninite formed at ca. 1,685 Ma suggesting that if the uraninite was unaltered, the PbO content in the uraninite would be near 20 wt.% and the UO_2 near 75 wt.% (Bowles 1990). The relatively high UO_2 , CaO, FeO and SiO_2 concentrations combined with the low PbO contents in the uraninite at Jabiluka (Table 6) indicate that all of the uraninite analysed in this study experienced post-precipitation alteration that removed Pb and introduced Ca, Fe and Si. Three populations of uraninite are revealed by the microprobe data and can be discriminated on a $\text{SiO}_2 + \text{CaO}$ concentration versus age diagram. One population forms a trend that intersects the *x*-axis at 1,690 Ma (Fig. 15). This age is indistinguishable from the oldest $^{207}\text{Pb}/^{206}\text{Pb}$ ages around 1,680 Ma given the scatter of the microprobe data. A second population forms a trend that passes through the *x*-axis at 1,385 Ma, similar to the $^{207}\text{Pb}/^{206}\text{Pb}$ of $1,302 \pm 37$ Ma and the U–Pb age 1358 ± 87 Ma obtained from Jabiluka uraninite. The third population does not define an obvious trend, but nevertheless is consistent with Pb loss and Ca and Si gain. Similar post-formation alteration of uraninite has been reported from unconformity-related uranium deposits in the Athabasca Basin, Canada (Kotzer and Kyser 1993; Fayek and Kyser 1997; Fayek et al. 2002) and Nabarlek (Polito et al. 2004).

Fig. 17 A simplistic, schematic representation of the formation of alteration styles surrounding the Jabiluka uranium deposit. **a** The Kombolgie Subgroup, deposited on the lower Cahill Formation shortly after 1,820 Ma, evolved during burial diagenesis to form diagenetic aquifers in the coarse-grained lithologies. Regional compression in the ARUF results in an increase in stress levels in the basement at ca. 1,680 Ma. Note the presence of the south-dipping contact at the unconformity. **b** Reverse-faulting at ca. 1,680 Ma immediately lowers the mean stress in the basement and forms breccias. Uraniferous basinal fluids are drawn into the basement resulting in the precipitation of uraninite and the alteration of the surrounding metamorphic assemblage. **c** The newly formed alteration minerals, having filled the available pore spaces, contribute to a mean increase in post-failure stress. Basinal fluids that were drawn into the basement during reverse faulting, but not consumed during alteration, equilibrate with the Mg- and Fe-bearing lithologies and are driven out of the ore zone during increased stress. This results in the formation of SC^1 chlorite in the overlying sandstone. **d** Post-failure stress continues to increase until the next rupturing event occurs and uraniferous basinal fluids are again drawn into the basement. Cross-cutting relations show that his process likely continued multiple times between 1,680 and 1,616 Ma. **e** Post-ore, reverse-faulting occurred after 1,616 Ma. U/Pb and $^{207}\text{Pb}/^{206}\text{Pb}$ ages suggest after 1,400 Ma. This faulting resulted in the formation of chlorite-quartz-dominated unmineralized breccias and brittle fault structures that truncate and displace mineralization at, or immediately below, the basement-sandstone contact. This style of faulting was later cut by normal faults (not shown in this orientation)

$^{40}\text{Ar}/^{39}\text{Ar}$ ages from SI^1 illite

Samples of SI^1 illite from diagenetic aquitards and aquifers in the Kombolgie Subgroup were analysed using the $^{40}\text{Ar}/^{39}\text{Ar}$ dating technique to compare the timing of fluid events in the Kombolgie Subgroup with the timing of uraninite formation in the Jabiluka deposit. The data and spectra for these samples are presented in Fig. 16 and Table 7.

Sample X198V 86.5 m contains minor amounts of illite in a SQ^1 dominated diagenetic aquitard. SI^1 illite extracted from this sample has an initial apparent age of 1,303 Ma, and a four-step plateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of $1,798 \pm 13$ Ma (Table 7; Fig. 16a). This age effectively records the time when diagenetic processes were terminated by the occlusion of pore fluids from the sample by early $\text{SQ}^{\text{a}/1}$ quartz cementation.

Petrological observations suggest that SI^1 illite from drill hole X198V 121.9 m formed in a diagenetic aquifer.

Table 7 Individual spot fusion ages, argon isotope ratios, % ^{40}Ar atmospheric contents and % ^{39}Ar released from SI¹ illite from Jabiluka

$^{36}\text{Ar}/^{40}\text{Ar}$	$^{39}\text{Ar}_k/^{40}\text{Ar}$	Ca/K	^{40}Ar atm. (%)	^{39}Ar (%)	$^{40}\text{Ar}/^{39}\text{K}$	Age (Ma)
X198-86.5m, SI ¹ illite: Plateau age = $1,798 \pm 13$ Ma						
0.000	0.027	0.010	6.17	3.72	34.92	$1,303 \pm 24$
0.000	0.017	0.014	1.69	14.20	56.80	$1,807 \pm 17$
0.000	0.018	0.014	0.36	20.82	55.57	$1,782 \pm 12$
0.000	0.018	0.015	0.12	29.69	56.84	$1,808 \pm 15$
0.000	0.018	0.014	0.07	24.96	56.72	$1,806 \pm 14$
0.000	0.019	0.013	0.00	6.32	53.95	$1,749 \pm 14$
0.000	0.033	0.063	0.01	0.21	30.42	$1,180 \pm 471$
0.000	0.032	0.119	0.02	0.07	31.66	$1,214 \pm 1156$
X198-121.9 m, SI ¹ illite: Plateau age = $1,747 \pm 11$ Ma						
0.000	0.021	0.016	13.14	4.13	41.87	$1,476 \pm 18$
0.000	0.016	0.022	6.79	3.16	56.87	$1,806 \pm 19$
0.000	0.018	0.024	2.33	11.71	54.13	$1,750 \pm 13$
0.000	0.018	0.019	0.94	8.12	54.42	$1,756 \pm 16$
0.000	0.018	0.024	0.46	26.17	53.87	$1,744 \pm 12$
0.000	0.018	0.021	0.21	23.36	54.19	$1,751 \pm 15$
0.000	0.019	0.022	0.10	19.64	53.47	$1,734 \pm 15$
0.000	0.026	0.017	0.31	3.43	38.84	$1,402 \pm 19$
0.003	0.188	0.414	84.32	0.28	0.81	44 ± 527
R129V-36.0 m, SI ¹ illite: Integrated age = $1,690 \pm 7$ Ma						
0.000	0.025	0.903	12.74000	2.09	34.5080	$1,265 \pm 6$
0.000	0.020	0.184	5.50000	1.98	46.7100	$1,560 \pm 7$
0.000	0.020	0.538	3.24000	2.29	49.2310	$1,616 \pm 6$
0.000	0.019	1.021	2.95000	8.30	50.9640	$1,653 \pm 6$
0.000	0.019	0.866	0.69000	9.65	52.1065	$1,678 \pm 8$
0.000	0.019	0.810	0.37000	4.84	52.62	$1,688 \pm 6$
0.000	0.019	1.159	0.36000	12.30	52.9670	$1,695 \pm 6$
0.000	0.019	0.469	0.25000	8.61	53.3690	$1,703 \pm 8$
0.000	0.019	0.184	0.04000	8.59	53.6790	$1,710 \pm 6$
0.000	0.019	0.892	0.12000	7.20	53.9210	$1,715 \pm 6$
0.000	0.018	0.443	0.29000	8.11	54.2700	$1,722 \pm 6$
0.000	0.018	0.197	0.25000	6.94	54.4710	$1,726 \pm 6$
0.000	0.018	0.288	0.20000	7.76	54.6990	$1,731 \pm 5$
0.000	0.018	0.791	0.06000	6.13	54.6590	$1,730 \pm 6$
0.000	0.019	1.013	0.08000	3.35	53.9390	$1,715 \pm 6$
0.000	0.020	1.159	0.19000	1.76	49.9080	$1,630 \pm 7$
N147V—130.1 m, SI ¹ illite; Plateau age = $1,683 \pm 11$ Ma						
0.001	0.034	0.012	17.97	1.47	24.02	$1,024 \pm 30$
0.000	0.022	0.006	10.10	1.96	40.79	$1,501 \pm 18$
0.000	0.021	0.019	4.15	6.47	46.04	$1,628 \pm 15$
0.000	0.021	0.022	1.25	8.86	46.73	$1,644 \pm 14$
0.000	0.021	0.020	0.71	27.56	48.16	$1,676 \pm 13$
0.000	0.021	0.017	0.30	25.20	48.37	$1,681 \pm 13$
0.000	0.020	0.023	0.22	18.09	48.89	$1,693 \pm 13$
0.000	0.021	0.020	0.30	10.40	48.47	$1,684 \pm 13$

This illite preserves a plateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of $1,747 \pm 11$ Ma (Fig. 16b), which corresponds to the 1,755 to 1,730 Ma Wonga Event in the Mt Isa Basin (Jackson et al. 2000) and the mid-Tawallah event in the southern McArthur Basin (Bull and Rogers 1996). Both of these events represent a period of basin inversion and erosion (Southgate et al. 2000). Sample R129V 36.0 m has similar petrological characteristics to sample X198 121.9 m but preserves a shallow, monotonic spectra that may be divided into two pseudo-plateau ages. The oldest recognizable six-step pseudo-plateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of $1,726 \pm 5$ Ma (Fig. 16c) may correspond to the waning stages of the Wonga Event, but it is also indistinguishable from the age of the Oenpelli Dolerite intrusion (Edgecombe et al. 2002; Kyser et al. 2000). A younger three-step pseudo-plateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of $1,688 \pm 7$ Ma

(Fig. 16c), is within error of the oldest $^{207}\text{Pb}/^{206}\text{Pb}$ ages of $1,679 \pm 17$ and $1,685 \pm 17$ Ma that correspond to original uraninite formation in the Jabiluka uranium deposit (Table 5), but the monotonic spectra suggests overprinting by a fluid at approximately 1,600 Ma. Another sample representing a diagenetic aquifer, N147V 130.1 m, has a four-step plateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of $1,683 \pm 11$ Ma (Fig. 16d), identical to the oldest $^{207}\text{Pb}/^{206}\text{Pb}$ ages from uraninite. Ages around 1680 Ma correspond to bend 2 on the Australian Apparent Polar Wander Path (Idnurm 2000), a period of uplift and erosion in the southern McArthur Basin and significant volcanic activity in the Mt Isa Basin (Page 1983; Southgate et al. 2000).

Discussion

Petrographic observations, $^{40}\text{Ar}/^{39}\text{Ar}$ data and stable isotope ratios from diagenetic minerals in the Kombolgie Subgroup overlying the Jabiluka deposit indicate that burial diagenesis, involving marine and minor meteoric fluids, had formed diagenetic aquitards and aquifers by 1,747 Ma (Fig. 16b; Kyser et al. 2000). An $^{40}\text{Ar}/^{39}\text{Ar}$ age from one diagenetic aquitard suggests that SQ^a and SQ^1 quartz cement formed and permanently terminated further diagenetic processes in some well-sorted sandstones as early as 1,798 Ma (Fig. 16a), indicating that the Kombolgie Subgroup is at least this old. U/Pb ages from the Plum Tree Creek Volcanics underlying the Kombolgie Subgroup indicate that sediments belonging to the Katherine River Group could have begun to accumulate shortly after $1,822 \pm 5$ Ma (Sweet et al. 1999a) and likely experienced shallow level diagenesis shortly after this. Thin section analysis and oxygen isotope data (Hiatt et al. 2005) show that porosity-occluding quartz cementation formed between 1.9 and 5 km, supporting our interpretation that burial diagenesis occurred during shallow burial shortly after deposition. Other $^{40}\text{Ar}/^{39}\text{Ar}$ ages indicate that diagenetic aquifers were well formed in the poorly sorted, fluvial clastic sediments by $1,747 \pm 11$ Ma (Figs. 16b, 17a). This is not a mineralization age, but corresponds to major tectonic events in the Mt Isa and southern McArthur Basins (Bull and Rogers 1996; Jackson et al. 2000) indicating that fluid migration and subsequent illite precipitation is sensitive to periods of tectonic instability. The six-step pseudoplateau $^{40}\text{Ar}/^{39}\text{Ar}$ age of $1,726 \pm 5$ Ma from one SI^1 illite sample may reflect the intrusion of the Oenpelli Dolerite into the ARUF (Kyser et al. 2000; Edgecombe et al. 2002), but it is equally plausible that this sample is recording Ar loss and overprinting of an earlier formed illite (e.g. $1,747 \pm 11$ Ma) by fluid flow during the formation of the Jabiluka deposit as indicated by five step ages between $1,695 \pm 6$ and $1,616 \pm 6$ Ma in the same sample (Fig. 16c).

LA-HR-ICPMS data reveal initial uraninite precipitation occurred in the basement at ca. $1,680 \pm 17$ Ma (Figs. 14, 17b). The oldest $^{207}\text{Pb}/^{206}\text{Pb}$ ages are coincident with an $^{40}\text{Ar}/^{39}\text{Ar}$ age of $1,683 \pm 11$ Ma from one SI^1 illite sampled above the deposit and numerous $^{40}\text{Ar}/^{39}\text{Ar}$ ages from illite extracted from diagenetic aquifers that occur locally and up to 70 km south and east of the deposit within Kombolgie Subgroup (Kyser et al. 2000). This suggests that fluids were moving out from the sandstone aquifers at this time to form the illite. Other $^{207}\text{Pb}/^{206}\text{Pb}$ ages to $1,596 \pm 16$ Ma likely represent minor Pb loss after initial uraninite precipitation, but they too coincide with more $^{40}\text{Ar}/^{39}\text{Ar}$ ages from diagenetic aquifers in the greater Kombolgie Subgroup suggesting that fluid flow out of these units continued until ca. 1,580 Ma (Kyser et al. 2000). Therefore it is possible that some uraninite was precipi-

itated at this late stage. The timing of uraninite and diagenetic illite precipitation coincides with at least three significant bends on the Australian apparent polar wander path at 1,680, 1,650 and 1,640 Ma (Idnurm 2000). Each of these bends: (1) represent changes in plate motion (Loutit et al. 1994; Idnurm and Giddings 1995), (2) would likely have reactivated faulting throughout the ARUF and (3) coincide with the formation of world class sediment-hosted Pb-Zn deposits in the southern McArthur and Mt Isa basins.

Mapping, petrographical observations and geochemical data show that finely disseminated to massive vein $\text{BU}^{1/2}$ uraninite precipitated with $\text{BC}^{1a/b}$ and BC^2 chlorite, $\text{BS}^{2/3}$ sericite, BQ^2 quartz and $\text{BH}^{2/3}$ hematite when a ca. 200°C, oxidised, basinal brine that evolved in the Kombolgie Subgroup aquifers was drawn into the basement after reverse-faulting lowered the mean stress at the sites of dilation (Fig. 17b). Mineralization appears to be best developed in association with breccia lenses that are broadly subconcordant to the $\text{S}_0/\text{S}_1/\text{S}_2$ foliation. In many cases, the syn-ore alteration minerals cross-cut one another and are found as brecciated fragments cemented by younger generations of uraninite, chlorite, sericite and hematite.

Fluid inclusion data from a syn-ore BQ^2 quartz– BU^1 uraninite vein indicates that the mineralizing brine was saline, but neither saturated, nor Ca-dominated. This brine could have originated from marine fluids, in contrast to the post-ore brines, which are Ca-dominated with Na:Ca ratios as low as 0.07 (Derome et al. 2003a). A marine origin for the uraniferous fluid is consistent with the stable isotope values from pre- and syn-ore phyllosilicates. Oxygen and hydrogen isotopic compositions of BS^1 sericite, $\text{BC}^{1a/b/2}$ chlorite and pristine uraninite grains indicate that a fluid with $\delta^{18}\text{O}_{\text{fluid}}$ and δD_{fluid} values of 4.0 ± 3.7 and $-27 \pm 17\text{‰}$, respectively (Fig. 13; Table 4), altered the metamorphic assemblage during mineralization. These values are indistinguishable from $\delta^{18}\text{O}_{\text{fluid}}$ and δD_{fluid} values obtained from SI^1 illite from diagenetic aquifers in the Kombolgie Subgroup overlying the deposit (Fig. 13; Table 4) and $\delta^{18}\text{O}_{\text{fluid}}$ and δD_{fluid} values obtained from illite from diagenetic aquifers up to 70 km south and east of the deposit throughout the Kombolgie Subgroup (Kyser et al. 2000). The $\delta^{18}\text{O}_{\text{fluid}}$ values from the pre- and syn-ore phyllosilicates in the sandstone and basement rocks are too low to be derived from a metamorphic fluid, whereas the high δD_{fluid} values rule out normal magmatic and pure meteoric sources (Taylor 1997), thus ruling out basement-derived fluids for the origin of the uranium. These minerals precipitated from basinal brines that evolved in sandstone aquifers from typically marine pore fluids with minor contributions from a meteoric source (Longstaffe 1987, 2000).

The fluid that formed SC^1 chlorite above the deposit cannot be isotopically distinguished from that associated with either the SI^1 illite or the basement-hosted alteration minerals. A likely scenario that adequately explains the occurrence of SC^1 chlorite in the illite-

dominated sediments, involves the stress cycling model of Sibson (1994). It is envisioned that basinal fluids are drawn into the basement after faulting and brecciation (Fig. 17b), thus precipitating uraninite and other alteration minerals and filling the available pore spaces. The basinal fluids equilibrate with the basement and those that did not partake in hydrating the basement assemblage are driven out of the ore zone during compression related to increased stress post-failure (Fig. 17c). This drives Mg-rich fluids into the overlying sandstone and results in the precipitation of SC^1 chlorite until the next rupturing event occurs and fluids are again drawn into the basement (Fig. 17d). Sibson (1994) notes that dilational fault jogs such as those described for Jabiluka are often “characterised by multiple re-cemented wall-rock breccias resulting from repeated hydraulic implosion”, exactly as observed at Jabiluka.

The presence of dismembered uraninite veins and discrete uraninite fragments within otherwise unmineralized breccia lenses, together with unmineralized breccias and brittle fault structures that truncate and displace mineralization at, or immediately below, the basement-sandstone contact (Fig. 17e) suggest that brittle deformation continued after original uranium precipitation ceased. Oxygen and hydrogen stable isotopes from BQ^4 quartz and BC^3/SC^2 chlorite indicate that evolved meteoric waters with low δD_{fluid} values entered the basement after mineralization ended (Fig. 13). Fluid inclusions in the BQ^4 quartz veins support the introduction of a different fluid from that, which deposited uranium. These post-ore quartz veins trapped fluids with low and high salinities and varied temperatures that suggest mixing occurred. (Fig. 12). The small population of high temperatures recorded by BC^3 chlorite and some fluid inclusions in the BQ^4 quartz veins may be representing the migration of fluids that were heated by the Maningkorri/Mudginberri phonolitic dykes or the more regional Derim Derim Dolerite. These fluids continued to evolve with regard to cation content. Fluid inclusions trapped at the tips of BQ^4 quartz veins and all inclusions trapped in the BQ^5/SQ^3 quartz veins have high salinities dominated by $CaCl_2$ (Fig. 12). These veins record low δD_{fluid} values typical of evolved meteoric water and are distinct from the mineralising brines (Fig. 13). Previous studies (Wilde et al. 1989; Derome et al. 2003a) have suggested that these low to highly saline $CaCl_2$ -dominated fluids represent mixing between two fluids with at least one of these fluids being the uraniferous brine. However, cross-cutting relations, stable isotopic data and new fluid inclusion evidence tied into a detailed paragenesis show that the $CaCl_2$ -dominated fluids are later than mineralization.

It has not been possible to determine exactly when each of these post-ore mineral-forming events occurred, but numerous $^{207}Pb/^{206}Pb$ ages between 552 and 1,596 Ma have been obtained (Table 5) with distinct populations occurring at ca. 802 ± 57 and $1,302 \pm 37$ Ma with a smaller population occurring at

$1,191 \pm 27$ Ma (Fig. 14). Sulfides, particularly galena (Fig. 10c, d), most likely formed during these post-mineralization events as radiogenic Pb was removed from uraninite. These ages correspond to regional tectonic events, including (1) the intrusion of phonolitic dykes between 1,370 and 1,316 Ma (Page et al. 1980; Needham 1988) and the intrusion of the Derim Derim Dolerite at $1,324 \pm 4$ Ma (Sweet et al. 1999b), (2) the amalgamation of Australia and Laurentia during the Grenville Orogeny at ca. 1,140 Ma (Wingate et al. 2002) and (3) the breakup of Rodinia between 1,000 Ma (Tack et al. 2001) and 750 Ma (Wingate and Giddings 2000). The youngest $^{207}Pb/^{206}Pb$ and chemical U/Pb ages between 384 and 715 Ma are recorded in uraninite that re-equilibrated with meteoric water having a $\delta^{18}O_{fluid}$ value between 0.3 and -13.9‰ . These values are much lower than those recorded by syn- and post-ore minerals, but the corresponding ages overlap the formation of the Cambrian to Early Ordovician Georgina Basin, which is partly overlain by Devonian freshwater sub-basins (Shergold and Druce 1980) and the extrusion of the 513 Ma Antrim Volcanics (Hanley and Wingate 2000). Sediments and volcanic flows related to each of these events cover more than 300,000 km² in northern Australia, suggesting significant tectonic activity and the potential for uraninite re-equilibration during this time.

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