

Formation and Evolution of the Centennial Unconformity-Related Uranium Deposit in the South-Central Athabasca Basin, Canada

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Abstract

The Centennial U deposit is situated in the south-central Athabasca Basin (Canada) and straddles the unconformity between early Paleoproterozoic to Archean metasedimentary and metavolcanic rocks and granitoids, and the clastic sediments of the Paleoproterozoic Athabasca Group. Although it has most characteristics of an unconformity-related uranium deposit, the Centennial deposit is atypical in that it is not directly associated with an electromagnetic conductor (there is a paucity of graphite in the basement) or with a major reverse fault zone; it is distal from a major fluid conduit (ca. 300 to 400 m from the Dufferin Lake Fault), has low Ni, Cu, Co, Zn, and Pb contents, and contains an unusually large amount (up to 5%) of secondary uranyl minerals. Additionally, a network of diabase dikes and sills is observed at Centennial, seemingly intruding the main U mineralization of massive uraninite, based on the relatively sharp contacts between the diabase dike and the high-grade U mineralization.

The pre-U alteration assemblage at Centennial includes kaolinite, illite, and sudoite, which have been formed by fluids with isotopic and chemical compositions that are comparable with those from other sandstone-hosted unconformity-type U deposits in the Athabasca Basin. Pre-U illite-related fluids have $\delta^{18}\text{O}$ of ca. 3‰ and δD of ca. –40‰, whereas pre-U chlorite-related fluids have $\delta^{18}\text{O}$ between 1.7 and 4.3‰ and δD between –18 and 1‰.

Laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS) U/Pb dating of the various U phases indicates that initial mineralization, represented by disseminated uraninite found directly to the north-northeast of the Centennial deposit, occurred at ca. 1.6 Ga. The main U mineralization, represented by massive and strongly altered uraninite, followed at an unknown time. A minor (<5%) unaltered uraninite formed from the local remobilization of the main massive uraninite at ca. 380 Ma. The main uranyl mineral, uranophane, formed last, at ca. 2 Ma. The recurrence of local U remobilization might have been facilitated by the persisting high permeability of the sandstones in the area due to the nearby Dufferin Lake Fault and to the emplacement of the diabase dikes.

The usefulness of Pb isotopes for exploration is demonstrated at the Centennial deposit, with strongly radiogenic Pb isotope signatures close to the deposit and a common Pb signature observed at a distance of a few km from the deposit.

Introduction

A RECENT discovery by Cameco Corporation (Jiricka, 2010), the Centennial deposit is situated in the south-central part of the Athabasca Basin, northern Saskatchewan, Canada (Fig. 1A, B). Exploration of the Virgin River trend started in 1959 with surficial geochemistry and geophysics, was reinitiated in 1992, and is currently operated by a joint venture between Cameco Corporation (49%), Areva Resources (49%), and Coronation Mines Ltd (2%). After identification of the major conductors by ground geophysics, the initial drilling covered the area to the northeast of the present deposit (Fig. 1C). The Centennial deposit proper was first intersected in 2004 (drill hole VR18), and subsequent drilling delineated the high-grade

U mineralization zone (Fig. 1D, E; Jiricka, 2010). The area immediately to the north-northeast of the Centennial deposit will be referred to as Virgin River-NNE, and the whole area studied is referred to as the Virgin River project.

In many respects, the Centennial deposit is a relatively typical example of unconformity-related uranium mineralization in the Athabasca Basin (Jiricka, 2010). The U mineralization is situated at the unconformity between the Athabasca Group sandstones and the underlying Paleoproterozoic basement (Fig. 1E). Mineralization-related alteration consists of silicification, desilicification, chlorite, kaolinite, dravite, and hematite in addition to disseminated, fracture-controlled, and massive U mineralization. However, the deposit is distinct in several of its characteristics (Jiricka and Witt, 2008), including the following: the mineralization is spatially unrelated with geophysical

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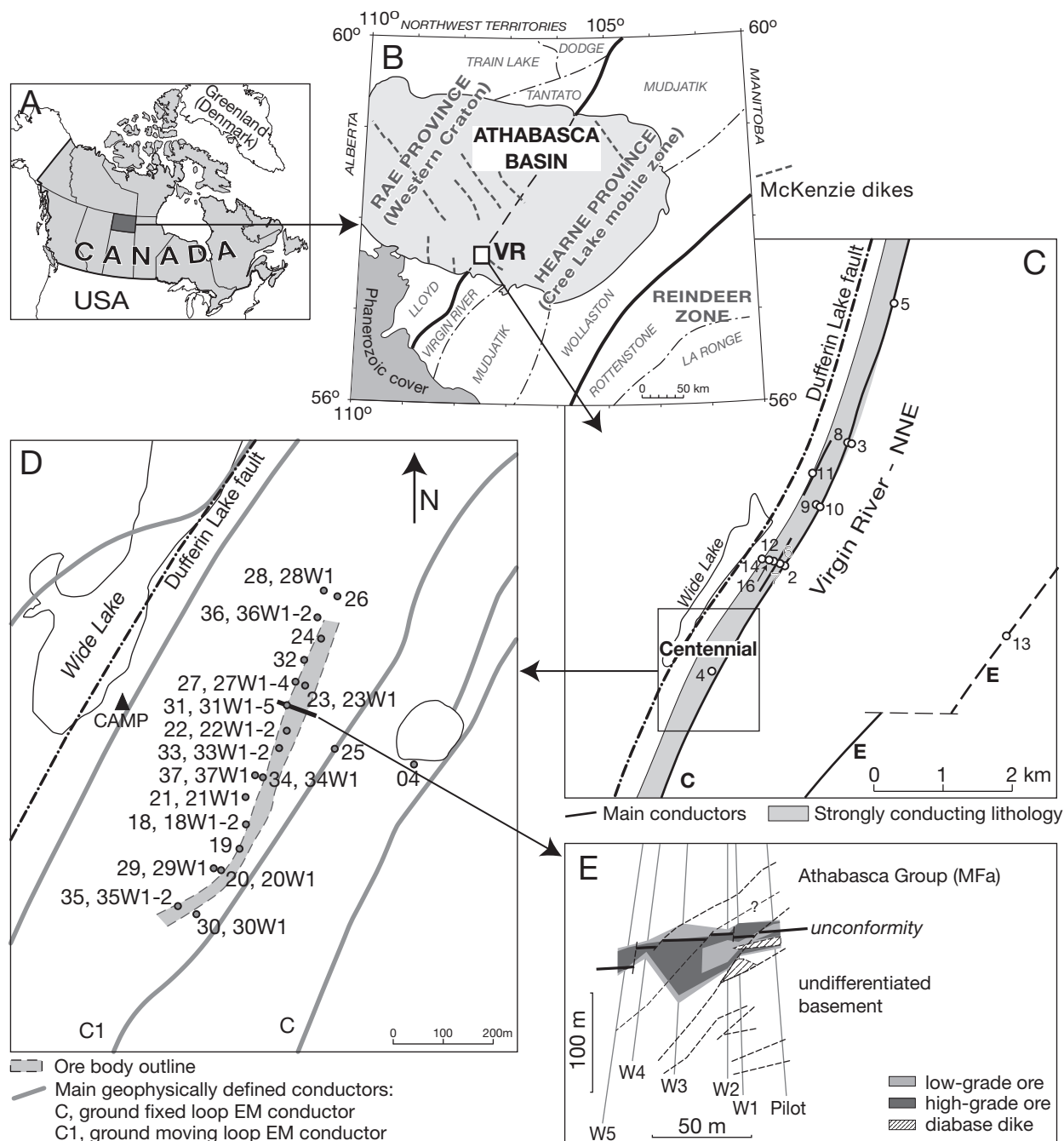


FIG. 1. General geologic setting of the Virgin River project (VR), situated in Canada (A) and in the Athabasca Basin (B). In C are shown the conductive lithology and the main conductors (defined by airborne geophysics), as well as the drill holes sampled in the Virgin River-NNE zone. The outline of the Centennial deposit, with its drill holes to date, is shown in D. An interpretive cross section, corresponding to fence VR31 (pilot hole and five wedges), is shown in E.

conductors; it is distal from the main fluid conduit (the Dufferin Lake Fault; Fig. 1D); there is a paucity of graphite in the basement (Reid et al., 2010); the deposit is not associated with a major reverse fault system, as are most deposits in the basin (Thomas et al., 2000); and the deposit has low Ni, Cu, Co, Zn, and Pb contents more akin to the orebodies of the McArthur River deposit (Thomas et al., 2000). An additional particularity of the Centennial deposit is the large amount of uranyl minerals, which are much more prevalent than in any other

Athabasca Basin deposit. Research on the Virgin River project has been minimal, and it is the goal of this paper to provide new petrological, geochemical, and geochronological data to understand the formation and evolution of this deposit and explain some of its distinguishing features.

Geologic Context

The Archean and Paleoproterozoic sub-Athabasca Group basement at the Virgin River project belongs to the Virgin

River Domain, which is the westernmost part of the Cree Lake mobile zone of the Hearne province (Fig. 1B). It is bounded to the west by the Lloyd Domain of the Rae province, and the contact is marked by the Virgin River shear zone. The exposed part of the shear zone, observed farther south, consists of a 4.5-km-wide zone of cataclases and mylonites. At Centennial, the Virgin River Domain is represented by metasedimentary and metavolcanic rocks, supracrustal schists and gneisses, felsic and mafic gneisses, and deformed Hudsonian-aged granitoids (Chandler, 1978). Metamorphic grade ranges from upper greenschist to lower amphibolite; retrograde greenschist facies is widespread. Intensive paleoweathering, consisting of kaolinite in the upper portions of the profile and illite and chlorite in the lower portions, affected the basement rocks prior to deposition of the Athabasca Group sediments. The Athabasca Basin initially formed at ca. 1750 Ma as a series of three NE-SW-oriented subbasins (Armstrong and Ramaekers, 1985; Kyser et al., 2000). Basin fill, up to 2 km thick, consists of thick clastic sequences of Paleoproterozoic age derived from the rapid exhumation of the Trans-Hudson orogenic belt (Ramaekers, 1990). At Centennial, the Athabasca Group is 800 to 900 m thick and consists entirely of the Manitou Falls Formation. The Manitou Falls-a member is represented by interbedded, matrix-supported quartz pebble conglomerate and poorly sorted medium-grained sandstone. The Manitou Falls-b member is dominated by medium-grained quartz sandstone with a substantial component of poorly sorted, clast-supported conglomeratic sandstone; however, it is only rarely present in the area. The Manitou Falls-c member is a relatively clean, medium- to coarse-grained sandstone with thin (<2 cm) granule and pebble beds. The Manitou Falls-d member is also a relatively clean, medium- to coarse-grained sandstone, but is characterized by the presence of a significant amount of clay intraclasts. The top 200 to 250 m of the Manitou Falls Formation is strongly silicified by quartz overgrowths, while moderate to strong desilicification is observed in the lower parts.

The major structural feature in the Virgin River project is the major NE-SW-oriented, NW-dipping, dextral oblique Dufferin Lake thrust fault. This structure has a strike length of more than 50 km and affects all rock types present (Fig. 1C), with a significant (ca. 250 m) post-Athabasca Group displacement (Reid et al., 2010) thrusting the western side up.

The Athabasca Group sedimentary rocks and the basement complex are intruded by a series of NW-trending mafic dikes belonging to the Mackenzie dike swarm (Cumming and Krstic, 1992; Fig. 1A). The dike trend corresponds to tensional directions associated with left-lateral movement along Hudsonian faults (Hoeve and Sibbald, 1978; Sibbald and Quirt, 1987). These intrusive rocks range from 1 m to several hundred meters in width and have a U-Pb age of 1267 ± 2 Ma (Le Cheminant and Heaman, 1989). At Centennial, several diabase dikes and sills are observed (Fig. 1E) and are probably related to the same magmatic event.

The Centennial deposit forms a NNE-SSW trend over a length of at least 650 m (Fig. 1D) and is found at a depth of approximately 800 m. The mineralization extends continuously from a few meters into the Athabasca Group to up to 30 m into basement (Fig. 1E; Jiricka and Witt, 2008). Several

high-grade intercepts have been described, with the most significant containing ca. 8.8 wt % U_3O_8 over 34 m (drill hole VR31-W3; Jiricka and Witt, 2008). Several small, lower-grade bodies of “perched” U mineralization are also present within the Athabasca Group, up to 100 m above unconformity.

Analytical Methods

Samples studied

Two hundred and fifty sandstone and basement samples were collected from most drill holes from the Virgin River project (Fig. 1C) and from a drill fence across the Centennial deposit (drill hole VR31, wedges 1–5; Fig. 1D, E). These samples represent all rock types, alteration styles, and mineralization types observed macroscopically. Polished thin sections were made from all samples and were used for petrological observations, using reflected and transmitted light microscopes and a scanning electron microscope (SEM), to establish the mineral paragenesis.

Electron microprobe analyses

The chemical compositions of alteration clay minerals and of U minerals were determined using electron microprobe analyses on the polished thin sections, using the 4 spectrometer Camebax MBX at Carleton University and the 5 spectrometer Cameca 100SX at the University of Manitoba. In both cases, the analytical conditions were as follows: acceleration voltage 15 KeV, beam current 20 nA, and beam size 2 mm. The background was measured on each side of the peak for half of the peak measuring time, which was between 20 and 50 s for the different elements. A suite of well-characterized natural and synthetic minerals and compounds was used for calibration standards.

Clay extraction

Alteration clay minerals were extracted from coarsely crushed samples (>1.4 mm) by ultrasound disintegration and were separated into size fractions by centrifugation (i.e., <2 μm , 2–5 μm , 5–10 μm , and >10 μm). All fractions were analyzed by X-ray diffraction (XRD) using a Philips (now known as Panalytical) X'pert Pro MPD system at Queen's University equipped with an X'celerator (high-speed) detector, copper target, 0.02° radian Söller slits on both incident and diffracted rays, a divergent slit of 0.5° , and an anti-scatter slit of 1° . The analytical conditions were as follows: step of 0.1° and counting time of 10 s. In all cases, the analysis was performed between 5 and $40^\circ 2\theta$, the accelerating voltage was 40 kV, and the current was 45 mA. Only pure, monomineralic separates (defined as having a minimum of 95% of one mineral, based on XRD results) were selected for subsequent stable isotope analyses.

Stable isotope analyses

For oxygen isotope analyses, approximately 5 mg of mineral separate was reacted with BrF_5 at ca. 550°C overnight. The extracted oxygen was converted to CO_2 in contact with a heated graphite rod, and the CO_2 was then analyzed offline using a Finnigan MAT 252 isotope ratio mass spectrometer (IRMS) at Queen's University (Canada). For hydrogen isotope analyses, encapsulated mineral separates were introduced in an automated TC/EA device, where hydrogen was

separated and then analyzed online using a DeltaplusXP IRMS at Queen's University, Canada. Replicate $\delta^{18}\text{O}$ analyses were reproducible to $\pm 0.2\text{‰}$, and δD values were reproducible to $\pm 3\text{‰}$. The results are given in the customary δ (δD and $\delta^{18}\text{O}$) notation, relative to Vienna standard mean ocean water (V-SMOW). The isotopic composition of H_2O -dominated fluids in equilibrium with the minerals analyzed was calculated using the fractionation factors of Eslinger and Savin (1973) for water-illite, Land and Dutton (1978) for water-kaolinite, and Yeh and Savin (1976) for water-chlorite. Hydrogen isotope fractionation factors used are those proposed by Yeh (1980) for water-illite, Gilg and Sheppard (1996) for water-kaolinite, and Graham et al. (1987) for water-chlorite.

Clay minerals formation temperature

The formation temperatures used for the fractionation calculations were estimated from the chemical composition of illite and chlorite and from literature data for kaolinite. For illite, a temperature of $180 \pm 10^\circ\text{C}$ was obtained using the molar fraction of pyrophyllite (20%, based on K, Al^{IV} , and Si^{IV} ; Cathelineau, 1988). This temperature is very similar to that obtained if the factor $\text{K} + (\text{Fe-Mg})$ in illite is considered (ca. $200 \pm 25^\circ\text{C}$; Battaglia, 2004). The temperature of 180°C is preferred, as the method is calibrated for similar temperatures and gives a better uncertainty margin. For chlorite, both oxide composition (SiO_2 , TiO_2 , and CaO , based on the data of Cathelineau and Nieva, 1985) and site occupancy (Al^{IV} ; Walshe, 1986; Cathelineau, 1988) were used, giving a temperature on the order of 200°C , with a fairly large uncertainty margin ($\pm 30^\circ\text{C}$). A similar temperature of ca. 190°C was obtained when the crystallinity of chlorite was considered (Arkai, 1991), but this temperature was not used, as there are uncertainties about the method calibration and XRD results variability. For kaolinite, neither mineral chemistry nor crystallinity can give reliable temperature estimates, so its formation temperature was estimated based on geologic factors, such as burial depth and literature data. Assuming a depth of 4 to 5 km during early diagenesis (e.g., Ramaekers, 1990) and a geothermal gradient of $30^\circ\text{C}/\text{km}$, the kaolinite formation temperature would be in the order of ca. 150°C , which is not dissimilar to what has been suggested for other parts of the basin (Alexandre et al., 2009; Cloutier et al., 2009).

U/Pb geochronology

U/Pb dating of U minerals was performed on polished thin sections by laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS) at Queen's University, Canada (Chipley et al., 2007). The laser used is the Mercantek® LUV213 Laser Ablation System with a frequency quintupled Nd-YAG (213 nm) used in spot mode. The laser conditions included laser power output of 30 to 40% ($<1 \text{ mJ}/\text{cm}^2$), pulses of 50 ms with repetition rate of 2 Hz, and spot size of 25 to 35 μm . The high resolution-inductively coupled plasma-mass spectrometry (HR-ICP-MS) instruments used are a double-focusing ThermoFinnigan Element2 used in medium resolution and in peak-jumping mode, and a multicollection ThermoFinnigan Neptune, equipped with nine Faraday cups and $10^{11} \Omega$ amplifiers for each.

Weak acid leaching

Weakly bound or adsorbed U and Pb were analyzed using the weak acid leaching method of Holk et al. (2003) at Queen's University (Canada). About 0.5 g of sample (size fraction 0.5–1.2 mm) and 5 ml of 0.02 M HNO_3 were loaded into a polyurethane tube, placed in an ultrasonic bath for 120 min, and then centrifuged. One g of the liquid was diluted to 50 g with ^{115}In -spiked 0.02 M HNO_3 acid, and the Pb isotope ratios and trace element concentrations were measured using a ThermoFinnigan Element2 HR-ICP-MS. The Pb isotope ratios were calculated using the signal intensities (counts/s) in low-resolution mode and have an average uncertainty of ca. 1%, based on repeat analyses. Corrections were made for procedure and analytical blanks, interferences from Hg, instrumental drift, and mass fractionation (Holk et al., 2003; Alexandre et al., 2009).

Results

Paragenetic sequence

Petrographic observations of Virgin River project samples using reflected and transmitted light and SEM (Fig. 2) indicate that the predominant detrital mineral present in the Manitou Falls Formation is quartz, with minor muscovite, tourmaline, and zircon. The quartz grains are usually rounded (Fig. 3A) and are typically less than 2 mm in diameter. As an expression of the early diagenetic stages, quartz precipitates as overgrowth rims, typically under 100 μm thick, on detrital quartz grains, underlain by a fine coating of hematite (indicating early oxidizing conditions; Fig. 3A), and locally resulting in complete disappearance of initial porosity in the sample. This silicification, which apparently occurred during early to peak diagenesis, is mostly observed in a layer situated at the base of the Manitou Falls-d member (between ca. 240 and ca. 310 m above the unconformity), where the initial porosity is entirely filled with quartz and very minor fine-grained kaolinite. This stage is followed by the formation of fine-grained kaolinite and of minor acicular, or spherulitic, dravite (Fig. 3A, B). Desilicification, manifest as the dissolution of the quartz overgrowth rims, occurs close to the unconformity and is accompanied by minor coarse-grained kaolinite (Fig. 3B). Near the uranium mineralization, late diagenesis pre-uraninite alteration is manifest as illite alteration of kaolinite, most often manifest by illite pseudomorphs of kaolinite (Fig. 3C). The relatively rare newly formed chlorite cutting across a kaolinite-illite mixture and the presence of chlorite pseudomorphs of kaolinite indicate that chlorite formed after kaolinite and illite (Fig. 3D). However, the relationship between illite and chlorite is less clear, as there is relatively little chlorite and their mutual crosscutting relationships are not always very well visible. Minor pyrite and rutile (indicating reducing conditions) seem to have formed after illite, as they form euhedral to subhedral crystals that crosscut illite. Minor euhedral quartz formed in mm-sized open fractures sometime after the pre-ore clay alteration, possibly after the early, disseminated uraninite.

At least four distinct types of U mineralization can be described at the Virgin River project. The first type, observed only in the Virgin River-NNE area but not at Centennial deposit sensu stricto, consists of disseminated uraninite grains not exceeding 0.5 mm across and precipitating in secondary

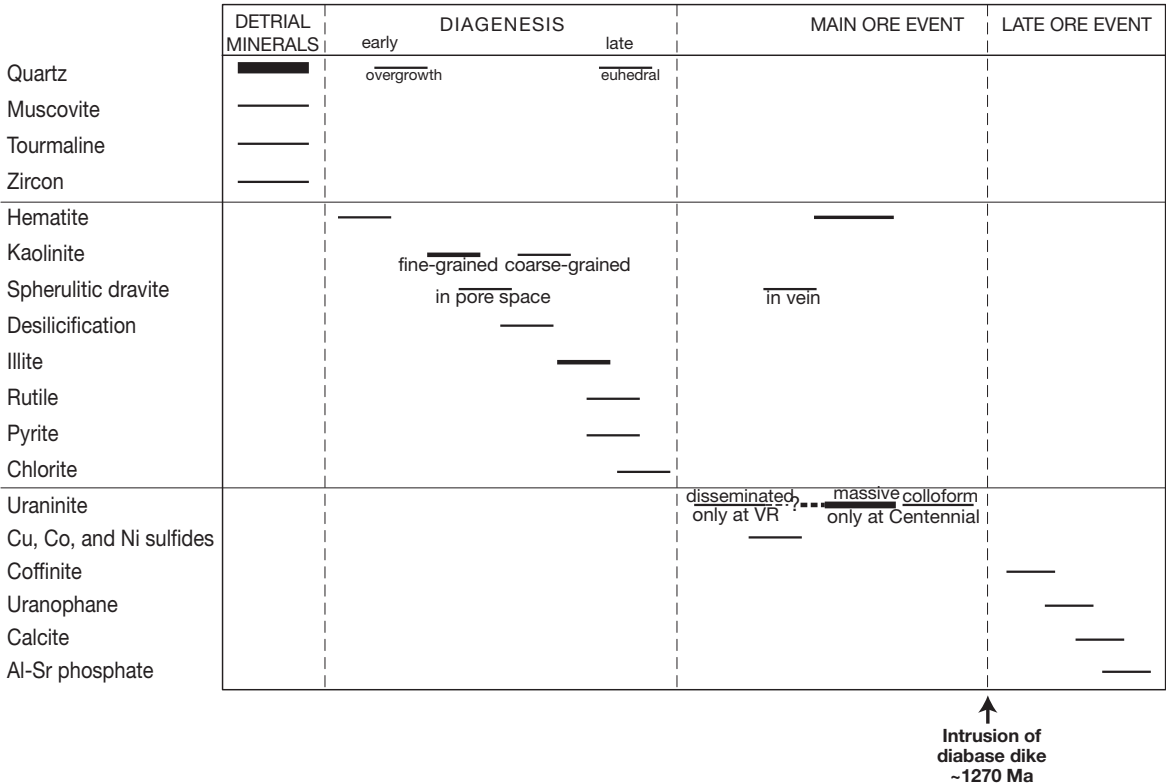


FIG. 2. Generalized paragenesis for the Virgin River project, combining observations from Virgin River-NNE and from the Centennial deposit.

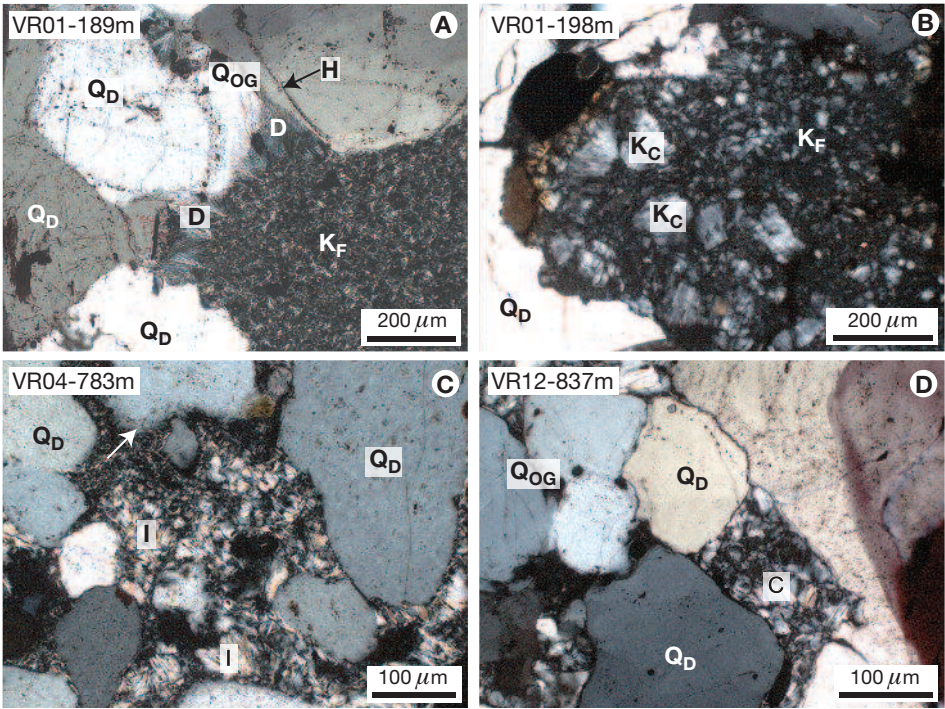


FIG. 3. Examples of pre-U clay alteration observed at Virgin River. A: moderate silicification by quartz overgrowth (QOG) over detrital quartz (QD) underlined by a thin hematite layer (H), with fine-grained kaolinite (K_F) and spherulitic dravite (D) forming in the pore space; B: coarse-grained kaolinite (K_C) forming after fine-grained kaolinite (K_F); C: illite (I) altering kaolinite and quartz dissolution (white arrow); and D: chlorite (C) alteration of kaolinite.

porosity between quartz, rutile, pyrite, and clay mineral grains (Fig. 4A), suggesting that it formed after these minerals. This uraninite is followed by narrow dravite veins, not exceeding 1 mm across, cutting across all these minerals. Ni, Co, and Cu sulfides formed next, indicating that reducing conditions were still prevalent and that the f_{S_2} was relatively high. The disseminated uraninite has been unevenly altered, as indicated by variability of its reflectance. Another type of U mineralization, forming the bulk of the ore at Centennial, consists of massive uraninite associated with hematite (Fig. 2). The presence of hematite indicates that oxidation of Fe^{2+} to Fe^{3+} was associated both temporally and spatially with reduction of

U^{6+} to U^{4+} and its precipitation as uraninite (e.g., Alexandre et al., 2005). The massive uraninite is almost invariably heavily altered, as indicated by zones of varying reflectivity, porosity, and a generally mottled appearance (Fig. 4B); only very small portions remain relatively fresh. The temporal relationship between the disseminated uraninite observed at Virgin River-NNE and the massive uraninite at Centennial is uncertain, as they were not observed in the same sample.

At some point, a network of diabase dikes and sills was emplaced at Centennial (Fig. 1E). The diabase dikes seem to intrude the main U mineralization of massive uraninite, based on the relatively sharp contacts between the diabase dike and

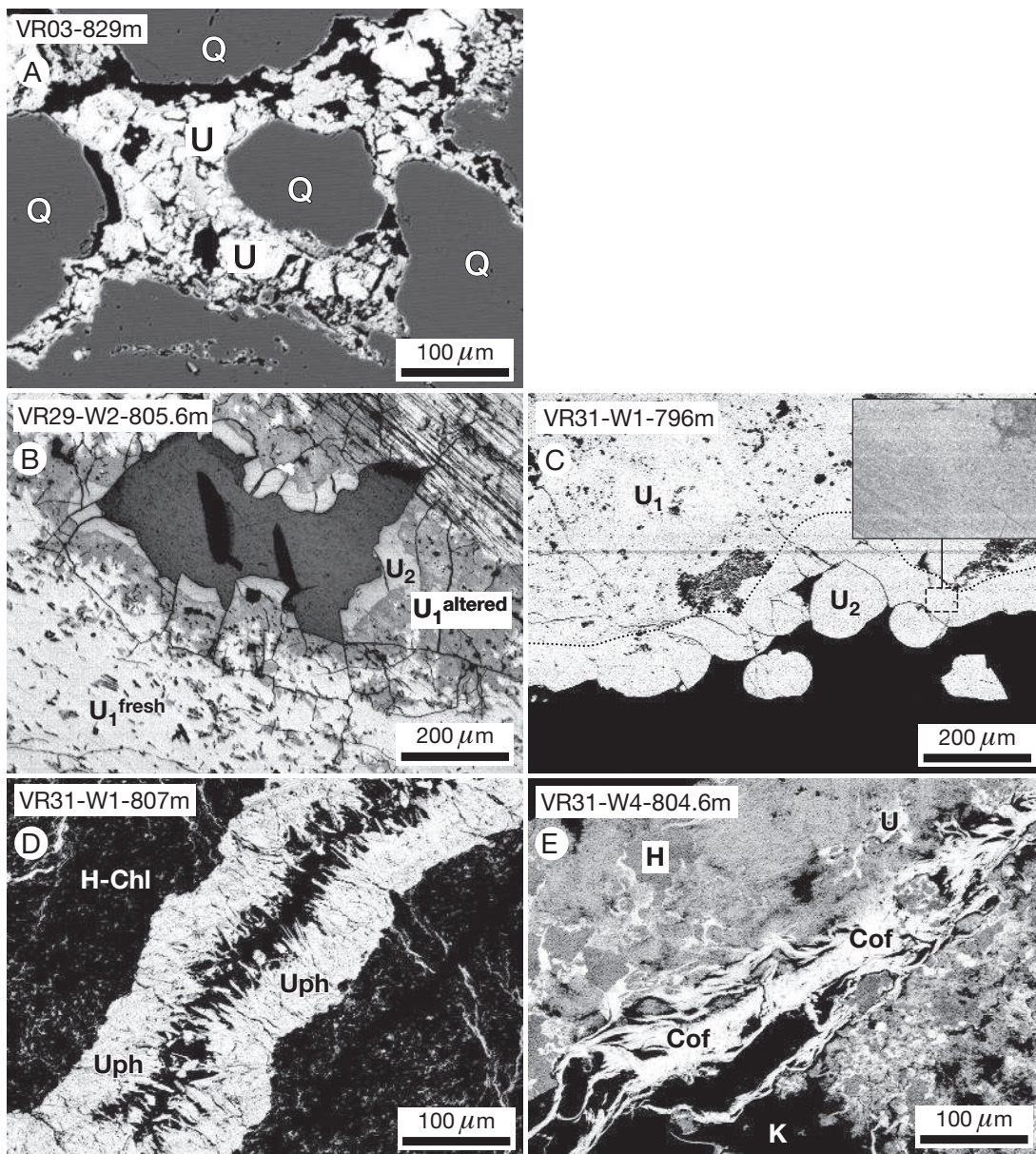


FIG. 4. U minerals observed at Virgin River-NNE and the Centennial deposit. A: disseminated uraninite (U) forming in the pore space between quartz grains (Q). B: visually fresh (U_1^{fresh}) and altered (U_1^{altered}) main stage massive uraninite, followed by late colloform uraninite (U_2). C: massive and strongly altered uraninite (U_1) followed by colloform and unaltered uraninite (U_2). Inset: U_2 structure. D: a relatively large uranophane (Uph) vein showing well-crystallized individual crystals and cutting through a mixture of hematite and chlorite (H-Chl). E: a vein of late fibrous coffinite (Cof) cutting through hematite (H) and uraninite (U).

the high-grade U mineralization (Reid et al., 2010). Plagioclase, one of the main constituents of the diabase, is heavily altered to sericite; chlorite alteration is also present. The diabase also contains opaque minerals, mostly rutile, which is sometimes strongly altered as well.

At Centennial, a distinct U mineralization type is represented by colloform uraninite (Fig. 4B), which differs from the earlier massive uraninite by its very low degree of alteration and by its preservation of initial textures. Colloform uraninite rarely exceeds 0.5 mm in size and formed on the surface of the earlier massive altered uraninite (Fig. 4B). Its presence suggests the persistence of reducing conditions at Centennial.

The last U mineralization type observed at Centennial is represented by small amounts of coffinite and uranophane (Fig. 2). Both form in fractures from 1 to 200 μm across and can form dense stockworks in hand sample and at the microscopic scale, particularly uranophane. Uranophane accounts for up to 5% of the mineralization at Centennial and often forms typical well-defined needle-shaped crystals (Fig. 4C), suggesting that it precipitated from a fluid rather than as an in situ alteration product of uraninite. The presence of a uranyl mineral such as uranophane at Centennial indicates a change to oxidizing conditions. Uranophane formed very late, as it crosscuts all other minerals, with the exception of some calcite and goyazite (Sr-Al phosphate).

The mineral paragenesis observed at the Virgin River project is comparable to that in other sandstone-hosted deposits in the Athabasca Basin (e.g., Thomas et al., 2000; Alexandre et al., 2009; Jefferson et al., 2007), with quartz overgrowths, a small amount of dravite, followed by illite and chlorite alteration associated with the U mineralization, minor sulfides, and predominantly disseminated uraninite. The atypical features of the Centennial U mineralization are the presence of strongly altered massive uraninite, later colloform uraninite dissociated from the main U stage, and a large amount of uranyl minerals, which have not been described elsewhere in the Athabasca Basin (e.g., Thomas et al., 2000).

Chemical composition of the alteration clay minerals

Kaolinite has a typical composition of ca. 46.7 wt % SiO_2 , ca. 38.6 wt % Al_2O_3 , and minor Fe, Mg, and Na (Table 1). The total is ca. 86.4 wt %, with the remainder of about 14 wt % being H_2O (Deer et al., 1992).

Illite from the Virgin River project has relatively high K (ca. 7.5 wt % K_2O) and is therefore not very interlayer deficient (K of ca. 0.9 apfu). In addition, the illite has relatively low degrees of octahedral site substitution (FeO and MgO of ca. 0.7 wt % each). Considering the interlayer deficiency and the octahedral site substitution, illite from the Virgin River project

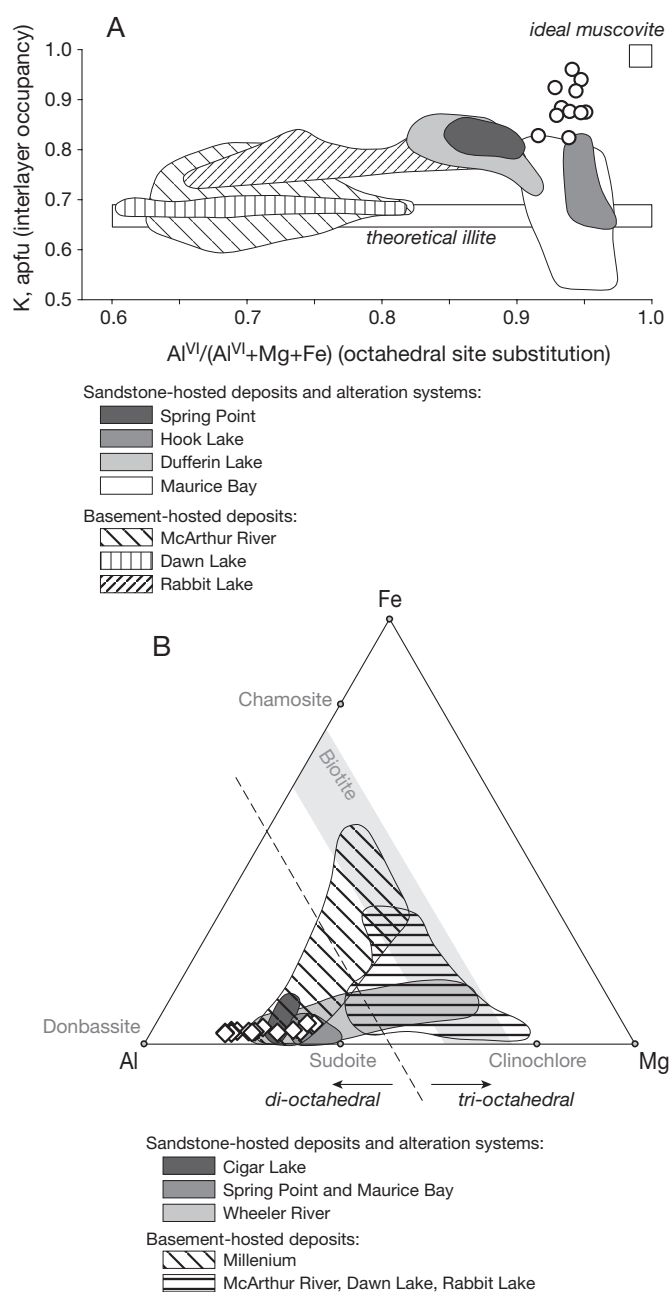


FIG. 5. Chemical compositions of illite (white circles: A) and chlorite (white diamonds: B) from the Virgin River project. Fields of the minerals from deposits elsewhere in the Athabasca Basin are given, as well as the theoretical position of main mineral species. Data from Alexandre et al. (2005, 2009) and Cloutier (2009).

TABLE 1. Average Chemical Composition (median values and standard deviation) of the Main Alteration Minerals Observed at the Virgin River Project

	SiO_2	Al_2O_3	FeO	MgO	CaO	Na_2O	K_2O	Total
Kaolinite (n = 11)	46.76 $\pm$ 2.12	38.63 $\pm$ 1.93	0.22 $\pm$ 0.17	0.10 $\pm$ 0.09	bdl	0.30 $\pm$ 0.11	bdl	86.37
Illite (n = 14)	51.63 $\pm$ 4.56	26.23 $\pm$ 1.74	0.73 $\pm$ 0.16	0.73 $\pm$ 0.15	0.25 $\pm$ 0.12	0.18 $\pm$ 0.1	7.51 $\pm$ 0.65	87.54
Chlorite (n = 15)	40.96 $\pm$ 4.30	32.80 $\pm$ 5.95	1.84 $\pm$ 0.84	6.00 $\pm$ 3.70	0.25 $\pm$ 0.11	0.12 $\pm$ 0.12	0.23 $\pm$ 0.34	84.92
Magnesiofite (n = 14)	36.24 $\pm$ 0.93	37.03 $\pm$ 2.41	1.13 $\pm$ 0.33	7.78 $\pm$ 0.59	0.13 $\pm$ 0.09	0.85 $\pm$ 0.16	bdl	83.53

Notes: Obtained by electron microprobe analysis; Ti, Cr, and Mn were analyzed as well, but were always below detection limit (bdl)

is very close to the ideal muscovite end member (Fig. 5A) and, given that it is relatively fine grained (Fig. 3C), it could be called sericite. When compared to illite compositions from other deposits in the Athabasca Basin, illite from the Virgin River project has the lowest degree of interlayer deficiency (Fig. 5A). The clearest differentiation between illites from basement-hosted and from sandstone-hosted systems seems to be the octahedral site substitution. From this point of view, illite from the Virgin River project is comparable to that of other sandstone-hosted systems in the basin, particularly the Maurice Bay deposit on the northwest margin of the basin (Fig. 5A; Alexandre et al., 2009). Additionally, it has been suggested that low interlayer deficiency and low degree of octahedral site substitution are typical for illite from Athabasca Basin ore zones (Quirt, 2010).

Chlorite from the Virgin River project has low Mg and Fe (MgO of ca. 6.0 wt % and FeO of ca. 1.8 wt %) and high Al (Al₂O₃ of ca. 32.8 wt %; Table 1), close to the ideal sudoite composition (Fig. 5B; Deer et al., 1992). Chlorites from sandstone-hosted deposits in the Athabasca Basin are typically dioctahedral (e.g., Percival and Kodama, 1989; Fig. 5B), with the exception of the Wheeler River chlorite, where the contribution from basement fluids may explain their partially trioctahedral character (Fig. 5B; Cloutier et al., 2009). Nevertheless, chlorites from the Virgin River project are typical of sandstone-hosted deposits in the Athabasca Basin (Fig. 5B).

Hydrothermal acicular tourmaline from the Virgin River project (Fig. 3A) contains less Mg and more Al (MgO ~7.8 wt %, Al₂O₃ ~37.0 wt %; Table 1) than typical dravite (MgO ~13.7 wt %, Al₂O₃ ~30.1 wt %; Deer et al., 1992). Compositionally, it is closer to magnesiofoitite (Hawthorne et al., 1999) than to dravite. Its total is ca. 83.5 wt %, due to the fact that B and H₂O were not measured; it would be close to 100% with the expected ca. 11 wt % of B₂O₃ and ca. 4 wt % of H₂O.

Stable isotopes

Diagenetic kaolinite at the Virgin River project has δD and $\delta^{18}O$ values of -85 to -52‰ and 15.6 to 16.9‰ , respectively (Table 2). Compared with other sandstone-hosted systems in the Athabasca Basin, the fluids in equilibrium with kaolinite from the Virgin River project have similar δD values but higher $\delta^{18}O$ values (Fig. 6).

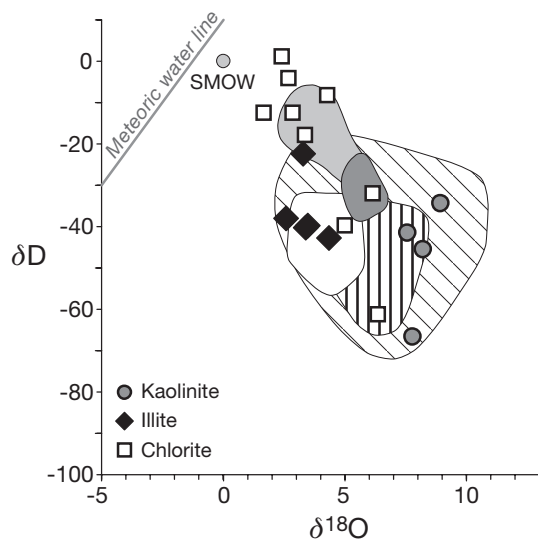
Pre-U illite from the Virgin River project has δD of -40 to -60‰ and $\delta^{18}O$ of 9.6 to 11.4‰ (Table 2). When the fluids in equilibrium with illite are compared with other deposits in the Athabasca Basin, they have slightly lower $\delta^{18}O$ (ca. 3‰) but similar δD values (ca. -40‰ ; Fig. 6).

Pre-U chlorites from the Virgin River project have a large variation in both O and H isotope composition, with δD of -100 to -43‰ and $\delta^{18}O$ of 7.0 to 11.7‰ (Table 2). The lowest δD values are probably caused by the recent interaction with modern meteoric fluids, which have δD values of ca. -130‰ , preferentially affecting the H composition but not the O values (Wilson and Kyser, 1987). When the least affected analyses are considered, the fluids in equilibrium with the Virgin River project chlorite vary between -18 and 1‰ for δD and between 1.7 and 4.3‰ for $\delta^{18}O$ (Table 2) and are slightly higher than fluids related to pre-U chlorites from other deposits in the Athabasca Basin (Fig. 6).

TABLE 2. Stable Isotopes Data for Alteration Clay Minerals at the Virgin River Project

Mineral	Mineral values				Fluid values	
	$\delta^{18}O$	Yield	δD	H ₂ O, wt %	$\delta^{18}O$	δD
Kaolinite						
VR12-501.5	16.9	15.6	-52	11.5	8.9	-34
VR12-250	15.6	16.6	-59	8.8	7.6	-41
VR12-908.5	15.8	16.1	-85	13.0	7.8	-67
VR12-734	16.2	16.4	-63	13.4	8.2	-45
Illite						
VR15-281	9.6	14.5	-56	3.9	2.6	-38
VR03-763	10.5	15.2	-57	5.7	3.5	-40
VR13-730	10.4	14.8	-58	4.7	3.4	-40
VR15-647.5	10.3	15.0	-40	4.1	3.3	-22
VR05-693.5	11.4	15.4	-60	4.5	4.3	-43
Chlorite						
VR04-599	8.7	15.3	-57	7.8	3.4	-18
VR04-698	8.0	15.6	-43	10.0	2.7	-4
VR04-775	9.6	15.9	-47	10.9	4.3	-8
VR04-793	8.1	16.2	-51	10.3	2.8	-13
VR4-807.5	11.5	16.6	-71	11.4	6.2	-32
VR04-821	11.7	16.1	-100	12.0	6.4	-61
VR12-145	10.3	15.7	-78	11.3	5.0	-40
VR04-792	7.0	15.0	-51	10.3	1.7	-13
VR04-789	7.7	14.7	-38	9.6	2.4	1

Notes: The mineral values and the values for fluid in equilibrium with the minerals are given; the temperatures used are 150°C for kaolinite, 180°C for illite, and 200°C for chlorite, based on mineral chemistry and literature data



Sandstone-hosted deposits systems (Maurice Bay and Spring Point):

■ illite
■ chlorite
□ kaolinite

Basement-hosted deposits (McArthur River, Dawn Lake, Rabbit Lake):

▨ illite
▨ chlorite

FIG. 6. H and O isotope composition of fluids in equilibrium with alteration clay minerals from the Virgin River project (circles, diamonds, and squares), with fields of the same minerals from other Athabasca Basin deposits. Data from Alexandre et al. (2005) and Alexandre et al. (2009).

U-Pb geochronology

The four discrete types of U mineralization described earlier (Figs. 2; 4) were analyzed by LA-ICP-MS (Table 3; Fig. 7).

Disseminated uraninite observed at Virgin River-NNE (Fig. 4A) shows strong radiogenic Pb loss, and all U-Pb analytical

points are situated close to the origin (Fig. 7A). However, two distinct discordia lines were defined, based on the $^{207}\text{Pb}/^{206}\text{Pb}$ ages obtained (Table 3), with upper intercepts at 1471 ± 37 and 1593 ± 47 Ma (Fig. 7A). These upper intercept ages are the same, within analytical uncertainty, as the corresponding average $^{207}\text{Pb}/^{206}\text{Pb}$ ages, 1437 ± 42 and 1593 ± 38 Ma,

TABLE 3. Results of the LA-ICP-MS Analyses Performed on Uranium Minerals in the Virgin River-NNE Area and the Centennial Deposit

Spot no.	Ratios						Ages					
	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm$	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm$
1. Virgin River, initial disseminated uraninite, low degree of alteration (sample VR03–827 m)												
1	0.1001	0.0035	0.0728	0.0025	0.0053	0.0002	1,625.0	65.0	71.0	3.0	34.0	1.3
2	0.0963	0.0034	0.0190	0.0007	0.0014	0.0001	1,553.0	62.0	19.0	1.0	9.0	0.6
3	0.0967	0.0034	0.0823	0.0029	0.0062	0.0002	1,562.0	62.0	80.0	3.0	40.0	1.3
4	0.1008	0.0035	0.0242	0.0008	0.0017	0.0001	1,638.0	66.0	24.0	1.0	11.0	0.6
5	0.0979	0.0034	0.0273	0.0010	0.0020	0.0001	1,585.0	63.0	27.0	1.0	13.0	0.7
6	0.0875	0.0031	0.0131	0.0005	0.0011	0.0000	1,371.0	55.0	13.0	1.0	7.0	0.1
7	0.0903	0.0032	0.0163	0.0006	0.0013	0.0000	1,432.0	57.0	16.0	1.0	8.0	0.1
9	0.0889	0.0031	0.0282	0.0010	0.0023	0.0001	1,401.0	56.0	28.0	1.0	15.0	0.7
10	0.0920	0.0032	0.0649	0.0023	0.0051	0.0002	1,468.0	59.0	64.0	3.0	33.0	1.3
11	0.0922	0.0032	0.0499	0.0017	0.0039	0.0001	1,472.0	59.0	49.0	2.0	25.0	0.6
12	0.0914	0.0032	0.0605	0.0021	0.0048	0.0002	1,455.0	58.0	60.0	2.0	31.0	1.3
13	0.0933	0.0033	0.0437	0.0015	0.0034	0.0001	1,493.0	60.0	43.0	2.0	22.0	0.6
14	0.0889	0.0031	0.0272	0.0010	0.0022	0.0001	1,402.0	56.0	27.0	1.0	14.0	0.6
2. Centennial, massive uraninite, various degrees of alteration (sample VR31-W1–807 m and VR31-W3–803 m)												
1	0.05325	0.00066	1.06849	0.01202	0.14464	0.00188	339.5	4.2	738.0	8.3	870.9	11.3
2	0.05286	0.00039	1.20640	0.01588	0.16191	0.00221	322.7	2.4	803.5	10.6	967.4	13.2
3	0.05402	0.00124	1.23129	0.02032	0.16536	0.00287	372.0	8.5	814.9	13.4	986.5	17.1
4	0.05197	0.00052	1.26243	0.01530	0.17276	0.00262	284.3	2.8	829.0	10.0	1,027.3	15.6
5	0.05273	0.00063	1.02247	0.00860	0.13814	0.00144	317.3	3.8	715.2	6.0	834.1	8.7
6	0.05258	0.00050	1.12562	0.01232	0.15242	0.00192	310.9	3.0	765.7	8.4	914.5	11.5
7	0.05277	0.00078	1.17942	0.01345	0.16081	0.00199	318.8	4.7	791.0	9.0	961.3	11.9
8	0.05197	0.00051	1.19292	0.01226	0.16496	0.00201	283.9	2.8	797.3	8.2	984.3	12.0
9	0.05280	0.00066	0.65209	0.02875	0.08952	0.00353	320.1	4.0	509.8	22.5	552.7	21.8
10	0.05267	0.00073	0.75939	0.03351	0.10490	0.00407	314.5	4.4	573.7	25.3	643.1	24.9
11	0.05277	0.00072	0.75110	0.03308	0.10365	0.00424	319.0	4.4	568.9	25.1	635.8	26.0
12	0.05272	0.00089	0.10946	0.00215	0.01498	0.00027	316.6	5.3	105.5	2.1	95.9	1.8
13	0.05253	0.00054	1.05663	0.00911	0.14468	0.00139	308.4	3.2	732.2	6.3	871.1	8.4
14	0.05255	0.00046	1.16725	0.01574	0.15852	0.00222	309.6	2.7	785.4	10.6	948.5	13.3
15	0.05231	0.00039	1.21809	0.01395	0.16718	0.00217	298.9	2.2	808.9	9.3	996.5	12.9
16	0.05286	0.00057	1.25899	0.01276	0.17067	0.00179	322.9	3.5	827.5	8.4	1,015.8	10.7
17	0.05362	0.00013	1.24246	0.01046	0.16867	0.00135	355.3	0.9	820.0	6.9	1,004.8	8.0
18	0.05357	0.00015	1.17380	0.01709	0.15913	0.00228	352.9	1.0	788.4	11.5	952.0	13.6
19	0.05361	0.00015	1.35898	0.00969	0.18378	0.00133	354.8	1.0	871.4	6.2	1,087.6	7.9
20	0.05355	0.00027	1.13505	0.01214	0.15409	0.00159	352.2	1.7	770.2	8.2	923.8	9.6
21	0.05357	0.00008	1.26878	0.01132	0.17176	0.00157	352.9	0.5	831.8	7.4	1,021.8	9.3
22	0.05358	0.00011	1.18970	0.00965	0.16098	0.00130	353.4	0.7	795.8	6.5	962.2	7.8
23	0.05363	0.00017	1.23686	0.01390	0.16795	0.00181	355.7	1.1	817.5	9.2	1,000.8	10.8
24	0.05360	0.00010	1.33720	0.00748	0.18099	0.00102	354.1	0.6	862.0	4.8	1,072.4	6.1
25	0.05356	0.00009	1.24178	0.01686	0.16821	0.00232	352.6	0.6	819.7	11.1	1,002.2	13.8
26	0.05360	0.00013	1.24647	0.01526	0.16789	0.00194	354.4	0.9	821.8	10.1	1,000.5	11.5
27	0.05358	0.00010	1.19372	0.00657	0.16147	0.00084	353.5	0.6	797.7	4.4	964.9	5.0
28	0.05346	0.00043	1.20354	0.02630	0.16357	0.00366	348.5	2.8	802.2	17.5	976.6	21.8
29	0.05351	0.00011	1.27950	0.01312	0.17273	0.00173	350.5	0.7	836.6	8.6	1,027.2	10.3
1	0.05353	0.00013	0.94070	0.00392	0.12748	0.00055	351.5	0.9	673.2	2.8	773.5	3.3
2	0.05343	0.00060	0.95738	0.01800	0.13002	0.00265	347.2	3.9	681.9	12.8	788.0	16.0
3	0.05353	0.00032	0.76741	0.00993	0.10406	0.00132	351.3	2.1	578.3	7.5	638.2	8.1
4	0.05357	0.00014	0.95800	0.00585	0.12976	0.00081	352.9	0.9	682.3	4.2	786.5	4.9
5	0.05360	0.00016	0.92754	0.00596	0.12552	0.00082	354.2	1.0	666.3	4.3	762.3	5.0
6	0.05340	0.00020	0.93391	0.00700	0.12684	0.00089	345.8	1.3	669.7	5.0	769.8	5.4
7	0.05313	0.00289	0.12816	0.00443	0.01747	0.00068	334.6	18.2	122.4	4.2	111.6	4.3
8	0.05323	0.00543	0.12131	0.00636	0.01656	0.00086	338.7	34.5	116.3	6.1	105.9	5.5
9	0.05308	0.00074	0.10610	0.00393	0.01451	0.00055	332.2	4.6	102.4	3.8	92.9	3.5
10	0.05277	0.00102	0.11147	0.00501	0.01534	0.00067	319.1	6.2	107.3	4.8	98.1	4.3
11	0.05290	0.00108	0.07971	0.00338	0.01095	0.00045	324.5	6.6	77.9	3.3	70.2	2.9
12	0.05352	0.00156	0.09174	0.00431	0.01244	0.00059	351.1	10.3	89.1	4.2	79.7	3.8

TABLE 3. (Cont.)

Spot no.	Ratios						Ages					
	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm$	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm$	$^{207}\text{Pb}/^{235}\text{U}$	$\pm$	$^{206}\text{Pb}/^{238}\text{U}$	$\pm$
3. Centennial, late colloform uraninite, no visible alteration (sample VR31-W1-796 m)												
1	0.05430	0.00168	0.02523	0.00108	0.00335	0.00014	383.3	16.4	25.3	1.1	21.6	0.9
2	0.05404	0.00031	0.17717	0.01626	0.02379	0.00213	372.7	34.2	165.6	15.2	151.6	13.6
3	0.05418	0.00031	0.13524	0.00691	0.01811	0.00090	378.5	19.3	128.8	6.6	115.7	5.8
4	0.05419	0.00091	0.02644	0.00081	0.00354	0.00008	378.9	11.5	26.5	0.8	22.8	0.5
5	0.05416	0.00010	0.20952	0.03102	0.02805	0.00415	377.8	55.9	193.1	28.6	178.3	26.4
6	0.05414	0.00012	0.18848	0.02240	0.02525	0.00299	376.8	44.8	175.3	20.8	160.8	19.0
7	0.05427	0.00033	0.35600	0.02307	0.04761	0.00284	382.1	24.8	309.2	20.0	299.9	17.9
8	0.05437	0.00038	0.35113	0.02017	0.04692	0.00245	386.3	22.2	305.6	17.6	295.6	15.5
9	0.05409	0.00023	0.47466	0.02062	0.06360	0.00284	375.0	16.3	394.4	17.1	397.5	17.7
10	0.05419	0.00021	0.45777	0.02192	0.06127	0.00293	378.9	18.1	382.7	18.3	383.3	18.3
11	0.05429	0.00007	0.24791	0.00534	0.03312	0.00072	383.2	8.3	224.9	4.8	210.0	4.6
12	0.05450	0.00028	0.28984	0.00497	0.03852	0.00055	391.7	6.7	258.4	4.4	243.7	3.5
13	0.05447	0.00030	0.23800	0.00618	0.03191	0.00077	390.4	10.1	216.8	5.6	202.5	4.9
14	0.05488	0.00538	0.33710	0.01298	0.04575	0.00162	407.4	15.7	295.0	11.4	288.4	10.2
15	0.05316	0.00088	0.20328	0.00831	0.02774	0.00130	335.8	13.7	187.9	7.7	176.4	8.3
16	0.05426	0.00046	0.21520	0.00430	0.02883	0.00061	381.9	7.6	197.9	4.0	183.2	3.9
4. Centennial, late vein uranophane (sample VR31-W1-807 m)												
2	0.04689	0.00367	0.00152	0.00017	0.00023	0.00000	43.4	3.4	1.5	0.2	1.5	0.0
3			0.00106	0.00010	0.00020	0.00000			1.1	0.1	1.3	0.0
10	0.05154	0.00335	0.00317	0.00017	0.00045	0.00002	265.2	17.2	3.2	0.2	2.9	0.1
13	0.05007	0.00242	0.00315	0.00012	0.00049	0.00001	198.1	9.6	3.2	0.1	3.1	0.0
14	0.05402	0.00675	0.00341	0.00036	0.00050	0.00001	371.7	46.5	3.5	0.4	3.2	0.1
15			0.00170	0.00014	0.00029	0.00001			1.7	0.1	1.8	0.0
16	0.05843	0.01371	0.00857	0.00080	0.01119	0.00010	546.1	128.2	8.7	0.8	7.7	0.6
17	0.04739	0.00267	0.00320	0.00013	0.00051	0.00001	68.8	3.9	3.2	0.1	3.3	0.1
18			0.00235	0.00015	0.00041	0.00001			2.4	0.2	2.6	0.1
19			0.00114	0.00010	0.00023	0.00000			1.2	0.1	1.5	0.0
22	0.05220	0.00382	0.01421	0.00020	0.00190	0.00005	294.4	21.6	14.3	0.2	12.3	0.3
23	0.05448	0.00459	0.01886	0.00025	0.00249	0.00010	390.9	32.9	19.0	0.3	16.0	0.6
			0.00138	0.00011	0.00023	0.00000			1.4	0.1	1.5	0.0
25	0.05195	0.00449	0.00434	0.00023	0.00062	0.00002	283.4	24.5	4.4	0.2	4.0	0.1
26	0.06161	0.00832	0.00411	0.00051	0.00051	0.00001	660.6	89.2	4.2	0.5	3.3	0.1
24	0.06032	0.00709	0.01500	0.00043	0.00169	0.00010	615.0	72.3	15.1	0.4	10.9	0.6
1	0.05380	0.00245	0.00228	0.00021	0.00031	0.00001	362.6	16.5	2.3	0.2	2.0	0.0
4	0.06167	0.00373	0.00472	0.00023	0.00055	0.00002	662.7	40.1	4.8	0.2	3.6	0.1
5	0.06427	0.00533	0.00880	0.00034	0.00098	0.00004	750.4	62.2	8.9	0.3	6.3	0.2
6	0.05196	0.00558	0.00244	0.00029	0.00032	0.00002	283.8	30.5	2.5	0.3	2.0	0.1
8	0.06468	0.00580	0.01161	0.00038	0.00127	0.00006	763.9	68.5	11.7	0.4	8.2	0.4
9	0.06053	0.00408	0.00649	0.00025	0.00075	0.00003	622.6	42.0	6.6	0.3	4.8	0.2
7	0.04864	0.00733	0.00148	0.00036	0.00021	0.00001	130.4	19.6	1.5	0.4	1.4	0.1

Note: The samples are organized in paragenetic order

respectively (Table 3). The age of 1593 Ma is the same as that of the main U deposition event proposed for the Athabasca Basin (Alexandre et al., 2007). The large analytical uncertainties for these ages reflect the strong loss of radiogenic Pb, causing highly discordant analyses (Fig. 7A).

The main stage massive uraninite is most often altered to a varying degree, with very few samples showing no visible alteration (Fig. 4B). The analyses of this apparently fresh uraninite correspond to a single discordia line, with an upper intercept of 752 ± 65 Ma and a lower one of 61 ± 24 Ma (Fig. 7B). The corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ ages average at ca. 640 Ma (Table 3).

All analyses of the variably altered main stage massive uraninite, which is the most abundant uranium phase at Centennial (Fig. 4B, C), form a single discordia line with an upper intercept at 350 ± 5 Ma, a lower intercept indistinguishable from the origin, and an average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 335 ± 21

Ma (Fig. 7C). However, the majority of the analyses are inversely discordant (plotting above the concordia line), whereas only a few are normally discordant (Fig. 7C). The inversely discordant points are from the most strongly altered uraninite, whereas the few normally discordant analyses correspond to the least altered zones of this uraninite. The inversely discordant data require either addition of Pb or loss of U. Lead is not compatible with the structure of uraninite and cannot be incorporated into its crystalline structure (e.g., Janeczek and Ewing, 1992; Finch and Murakami, 1999). If excess Pb is present within the areas analyzed, it can only be present as distinct mineral domains such as micron-size galena grains. However, larger grains of galena were observed in this sample, suggesting that Pb is not present as galena microinclusions in uraninite and therefore supporting the possibility that U has left uraninite in response to the strong alteration observed (Fig. 4B, C).

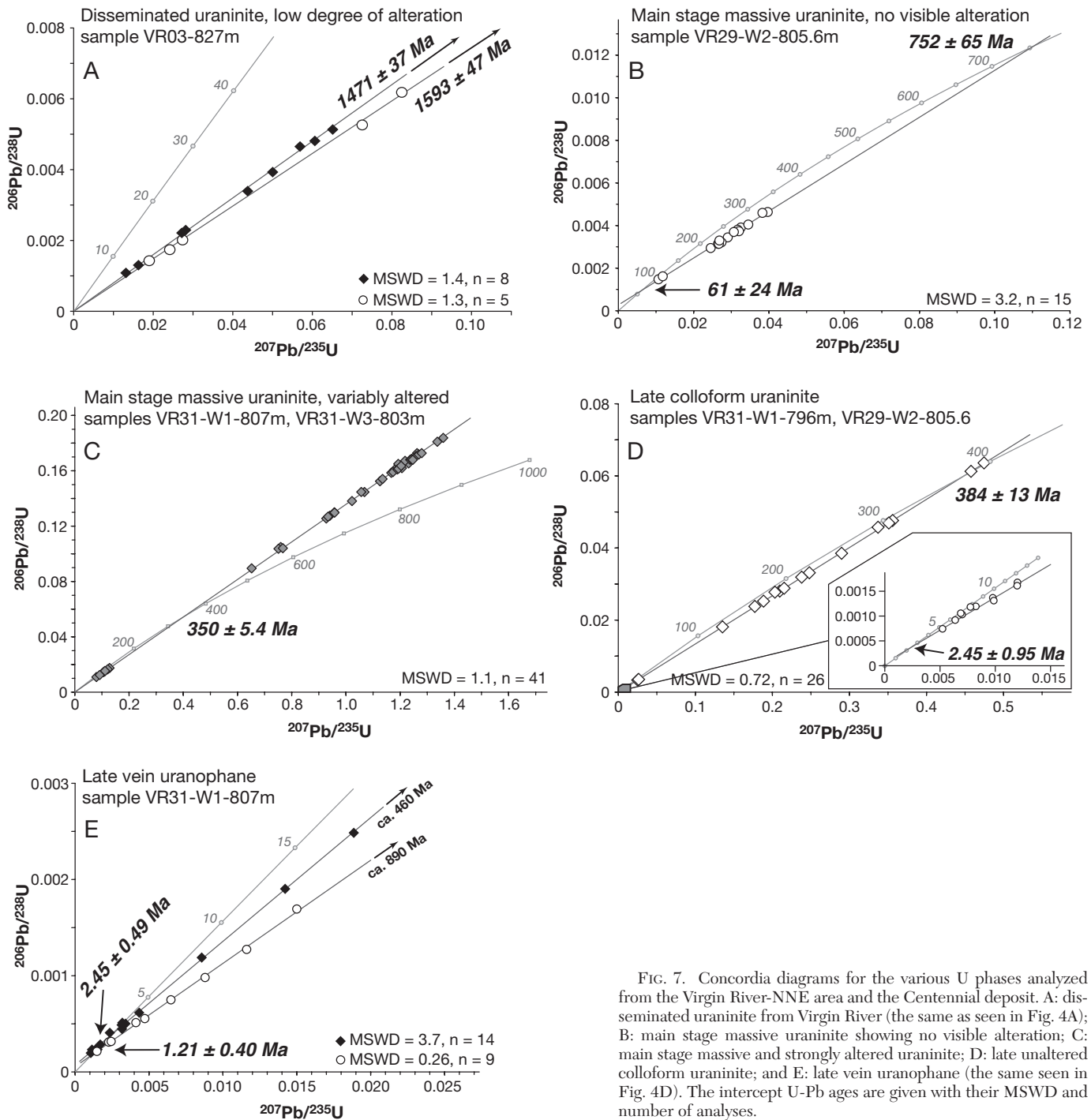


FIG. 7. Concordia diagrams for the various U phases analyzed from the Virgin River-NNE area and the Centennial deposit. A: disseminated uraninite from Virgin River (the same as seen in Fig. 4A); B: main stage massive uraninite showing no visible alteration; C: main stage massive and strongly altered uraninite; D: late unaltered colloform uraninite; and E: late vein uranophane (the same seen in Fig. 4D). The intercept U-Pb ages are given with their MSWD and number of analyses.

The late colloform and alteration-free uraninite observed at Centennial (Fig. 4B, C) has U-Pb isotope compositions that define a single and well-defined discordia line with upper intercept at 384 ± 13 Ma and a lower intercept of 2.45 ± 0.95 Ma (Fig. 7D). The corresponding $^{207}\text{Pb}/^{206}\text{Pb}$ ages vary between ca. 50 and ca. 400 Ma, with most at ca. 350 Ma. The absence of visible alteration in the colloform uraninite (Fig. 4C) and the loss of radiogenic Pb during the recent event would suggest that the upper intercept corresponds to its initial formation.

Analyses of well-crystallized uranophane (Fig. 4D) give two discordia lines with imprecise upper intercept ages (ca. 460 and ca. 890 Ma) and lower intercept ages at 2.45 ± 0.49 and 1.21 ± 0.40 Ma (Fig. 7E). Because of the very low amount of Pb, the $^{207}\text{Pb}/^{206}\text{Pb}$ ages are imprecise and highly variable, ranging between ca. 50 and ca. 750 Ma (Table 3). Most of the U-Pb analytical points are near or at the lower intercept (Fig. 7D), underlining the uncertainty of the upper intercept ages.

Weak acid leaching

The U concentrations and the Pb isotope compositions of the weakly bound and adsorbed fraction analyzed by weak acid leaching vary as a function of proximity to the unconformity and to the deposit (Fig. 8, Table 4). Near the present-day surface, leachable U concentrations are often below 1 ppb, and the $^{207}\text{Pb}/^{206}\text{Pb}$ ratios vary between ca. 0.5 and ca. 0.8 (Table 4, Fig. 8). At or near the unconformity, leachable U exceeds 10^4 ppb and the $^{207}\text{Pb}/^{206}\text{Pb}$ ratio drops below 0.1 (Table 4). Samples from the northernmost drill hole, situated ca. 6 km to the north of the Centennial deposit (VR05; Fig. 1C), have U concentrations slightly above the detection limit and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of ca. 0.7 to 0.8 (typical of common Pb), irrespective of depth (Table 4). U concentration is inversely correlated with the $^{207}\text{Pb}/^{206}\text{Pb}$ ratio, with analyses situated on a continuum between a common Pb component ($^{207}\text{Pb}/^{206}\text{Pb}$ ratio of ca. 0.8) and a radiogenic Pb component ($^{207}\text{Pb}/^{206}\text{Pb}$ ratio of ca. 0.1 associated with high U concentrations; Fig. 9). This inverse correlation (or positive correlation, if the $^{207}\text{Pb}/^{204}\text{Pb}$ or $^{206}\text{Pb}/^{204}\text{Pb}$ ratios are plotted against U) indicates that the weakly bound Pb analyzed originates from the disintegration of U. Variations around the mixing curve (Fig. 9) underline the difference in mobility of U and Pb. The $^{207}\text{Pb}/^{206}\text{Pb}$ ratio of ca. 0.1 associated with the samples with the highest leachable U concentrations corresponds to an age of ca. 1.6 Ga, which is the age for the formation of the majority of the unconformity-related deposits in the Athabasca Basin (Alexandre et al., 2007).

Discussion

Timing of the mineralization events

The disseminated and relatively low grade uraninite mineralization, observed only at Virgin River-NNE (Fig. 4A), was

dated at 1593 ± 47 Ma using U-Pb isotope compositions, confirmed by the corresponding average $^{207}\text{Pb}/^{206}\text{Pb}$ age (Fig. 7A). This age is the same as the one suggested for the majority of unconformity-related U deposits in the Athabasca Basin (Alexandre et al., 2007), indicating that the Virgin River area was involved in this basin-wide U deposition event.

The visibly unaltered main stage uraninite (Fig. 4B) gives an age of 752 ± 65 Ma, whereas the porous and variably altered massive uraninite (Fig. 4C) has an age of 350 ± 5 Ma (Fig. 7C). Both ages are much lower than what has been suggested for any deposit in the Athabasca Basin (Baadsgard et al., 1984; Carl et al., 1992; Cumming and Krstic, 1992; Philippe et al., 1993; Alexandre et al., 2007). The late colloform uraninite formed at 384 ± 13 Ma (Fig. 7D). Petrographically, it postdates the main stage massive uraninite (Fig. 4B, C), probably using some of the U removed from the massive uraninite, as suggested by the porous quality of the latter and its inverse discordia U-Pb analyses (Fig. 7C). This discrepancy suggests that the age of ca. 350 Ma does not correspond to an actual geologic event (e.g., a period of fluid circulation), but rather to the combined effects of (1) repeated disturbances to the U-Pb isotope system of this uraninite, and (2) its high porosity (Fig. 4C), which might have made it strongly reactive with fluids even after the formation of the colloform uraninite.

It could therefore be concluded that the ca. 350 Ma date can most probably be rejected as the initial formation age of the main stage massive uraninite at Centennial. This would seemingly mean that this uraninite formed at 752 ± 65 Ma, which is the only other age directly obtained on it (Fig. 7). Another possibility is that the main stage uraninite formed much earlier, possibly as early as the disseminated uraninite at Virgin River-NNE (1593 ± 47 Ma) and many deposits in

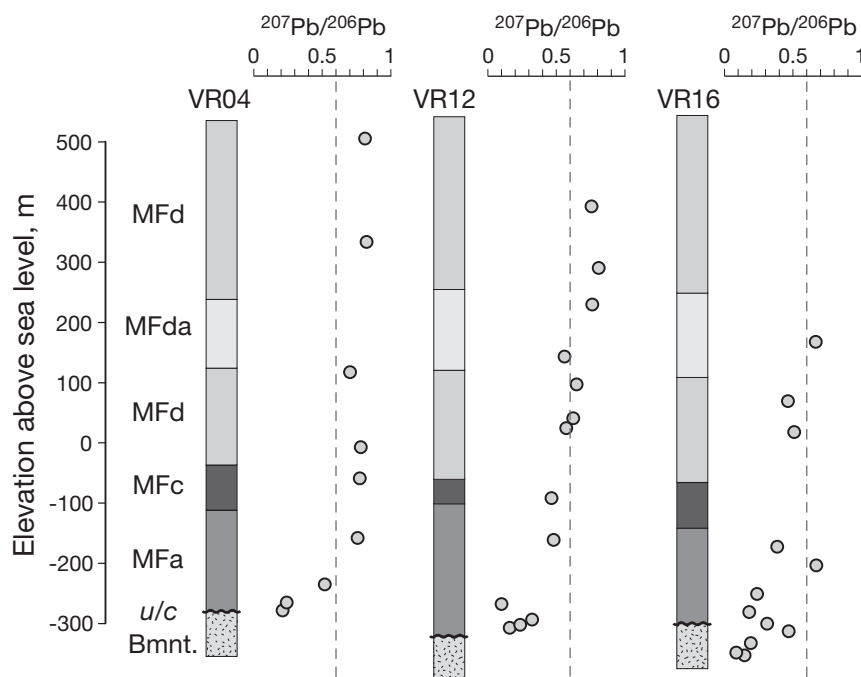
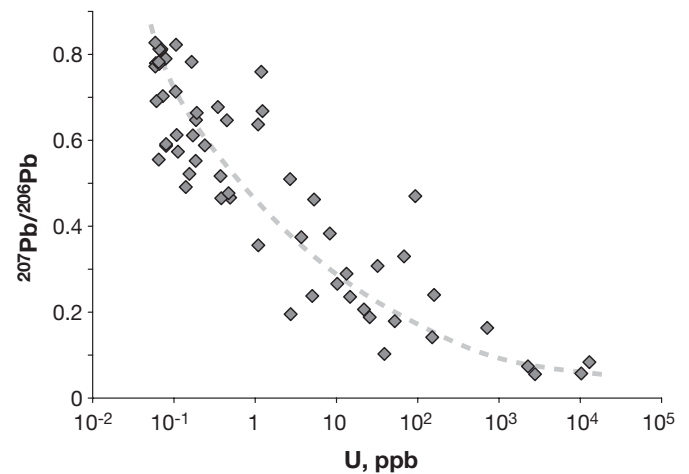


FIG. 8. $^{207}\text{Pb}/^{206}\text{Pb}$ ratios in the weakly bound phases, obtained by weak acid leaching, plotted as function of depth for selected drill holes (drill hole locations are shown in Fig. 1C).

TABLE 4. U Concentration and Pb Isotope Composition in the Leachable Phase in Samples from the Virgin River Project

Drill hole	Sample depth (m)	U (ppb)	²⁰⁶ Pb/ ²⁰⁴ Pb	²⁰⁷ Pb/ ²⁰⁴ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb
VR03	428	0.07	29.97	16.64	0.5552
	763	0.19			0.6472
	791	0.08	19.34	15.28	0.7900
	834	10,278.29	4,634.29	265.22	0.0572
	835	2,770.13			0.0557
	839	2,268.78	770.11	56.90	0.0739
	870	0.16	32.18	16.79	0.5218
VR04	33	0.07	18.38	14.87	0.8093
	206	0.11	19.39	15.95	0.8224
	422	0.07	19.06	13.39	0.7028
	546	0.06	19.61	15.29	0.7796
	599	0.06	17.53	13.53	0.7719
	698	1.19	75.18	57.11	0.7595
	775	0.37	27.79	14.36	0.5168
	808	14.71	100.50	23.68	0.2356
	821	21.78	99.13	20.45	0.2063
VR05	131	0.07	20.03	16.28	0.8125
	175	0.07	19.56	15.89	0.8123
	694	0.06			0.6912
	761	0.17	29.37	22.99	0.7826
VR06	796	0.11	22.40	13.72	0.6124
	853	1.09	58.38	20.76	0.3557
	868	10.22	93.74	24.95	0.2661
VR07	816	0.49	37.40	17.45	0.4666
	839	0.38	36.53	16.99	0.4652
VR11	827	2.73	101.58	19.84	0.1953
	831	13.30	72.24	20.92	0.2896
	839	3.69	57.26	21.44	0.3745
VR12	145	0.07	20.63	16.05	0.7779
	250	0.06	15.86	13.12	0.8273
	312	0.07	18.69	14.63	0.7825
	380	0.11	30.32	17.39	0.5735
	445	0.19	21.86	14.51	0.6640
	502	1.09	34.11	21.73	0.6370
	516	0.24	33.67	19.81	0.5883
	632	0.47	115.69	55.17	0.4769
	702	0.14			0.4912
	809	38.85	382.10	39.27	0.1028
	837	67.82	59.76	19.71	0.3298
	846	158.65	93.03	22.34	0.2401
	850	714.79	144.04	23.54	0.1634
VR13	652	0.19	24.60	13.59	0.5522
	750	0.45			0.6467
	827	0.08	103.37	60.69	0.5871
VR15	281	0.11			0.7133
	648	0.08			0.5911
	810	0.17	82.34	50.34	0.6114
VR16	375	1.23	24.75	16.53	0.6681
	473	5.30	37.33	17.25	0.4622
	525	2.69	33.45	17.05	0.5098
	715	8.27	46.04	17.64	0.3830
	745	0.35	23.42	15.86	0.6775
	795	5.03	83.42	19.82	0.2377
	825	52.12	135.27	24.21	0.1790
	844	32.02	64.74	19.92	0.3077
	855	93.32	35.47	16.68	0.4701
	876	25.61	122.77	23.11	0.1883
	892	12,929.14	936.97	78.40	0.0837
	896	150.58	210.11	29.74	0.1415

Weak acid leaching-ICP-MS analyses

FIG. 9. ²⁰⁷Pb/²⁰⁶Pb ratio plotted as a function of U concentration in the weakly bound phases (weak acid leaching-ICP-MS), showing a mixture between common Pb, derived from low U sources, and highly radiogenic Pb, derived from sources with high U.

the Athabasca Basin (Alexandre et al., 2007), but was subject to a total resetting at ca. 752 Ma. In support of this hypothesis is the fact that the main stage massive uraninite ore is crosscut by the diabase dikes (Reid et al., 2010), which most probably belong to the McKenzie igneous event, dated at 1267 ± 2 Ma (Le Cheminant and Heaman, 1989). There are other arguments suggesting that the age of ca. 752 Ma represents a resetting rather than a formation age for this uraninite: (1) this age has never been recorded as the formation age for the main stage ore for any deposit in the Athabasca Basin, and it has never been suggested for a main stage U ore in any U deposit in the Athabasca Basin (Baadsgard et al., 1984; Carl et al., 1992; Cumming and Kristic, 1992; Philippe et al., 1993); (2) it has been observed in several U deposits in the Athabasca Basin only as a perturbation age due to far-field tectonic events (e.g., Alexandre et al., 2007); (3) by the Neoproterozoic, the Athabasca Basin was well beyond its maximum depth and peak diagenesis, leading to lower temperatures (Kyser et al., 2000) and probably to conditions less suitable for the formation of a major U orebody. Considering that uraninite is a chemically active mineral and readily exchanges elements or recrystallizes during fluid circulation events, it is not implausible to suggest that the age of ca. 752 Ma obtained for the main stage uraninite at Centennial corresponds to such an event related to the breakup of Rodinia rather than to the initial formation age for this phase.

Therefore, the initial crystallization age for the main stage massive uraninite at Centennial is probably between 1267 (McKenzie dikes) and 1593 Ma (disseminated uraninite at Virgin River-NNE). It is even plausible to surmise that main stage uraninite formed closer to ca. 1593 Ma, as U deposition is documented in the area at that time (Fig. 7A), as it is elsewhere in the Athabasca Basin (Alexandre et al., 2007). In this case, the main stage massive altered uraninite at Centennial would have been subjected to several events that disturbed its U-Pb isotope system, removing U and Pb in different proportions. These events include the following: the emplacement of the McKenzie dikes at ca. 1267 Ma (Le Cheminant and Heaman,

1989), the breakup of Rodinia, reflected by the 752 ± 65 Ma age, the fluid event leading to the formation of the colloform uraninite (ca. 380 Ma; Fig. 7D), possibly Alpine tectonic activity in western Canada (the lower intercept age of 61 ± 24 Ma for the unaltered main stage uraninite), the recent formation of uranyl minerals, and recent Pb loss (Fig. 7C).

The presence of such an array of ages indicates that the Virgin River system remained open to fluid circulation over most of its history and underlines the distinctiveness of the Centennial deposit in terms of fluid flow. Normally, such active fluid circulation would be detrimental to the preservation of U mineralization. For the Centennial deposit to be preserved, the fluids involved would have had to be reducing and poor in complexing elements such as P, C, F, and Cl, thus minimizing their ability to mobilize U. As a corollary, the whole Virgin River area would have had to be relatively well shielded from the more oxygen rich upper parts of the Athabasca Basin.

The late, unaltered colloform uraninite gives a single discordia line with a well-defined upper intercept at 384 ± 13 Ma (Fig. 7D). Given that there is no indication of any alteration, recrystallization, or remobilization of this uraninite (Fig. 4B, C), it was likely affected only by radiogenic Pb loss, as seen by the distribution of the Pb/U ratios (Fig. 7D, Table 3); therefore, the age obtained can be considered the formation age for this uraninite. Even though this is a relatively volumetrically minor mineralization stage of the Centennial deposit, such an age is highly unusual for U ore in the Athabasca Basin and has rarely been observed here. For example, in one of the few documented cases, the late uraninite from the Eagle Point deposit, situated on the northeast edge of the Athabasca Basin, gives a well-defined upper U-Pb intercept age of 403 ± 21 Ma (Cloutier, 2009). Thus, a significant fluid circulation event occurred at ca. 380 Ma in the south-central Athabasca Basin, strongly affecting the isotopic system of the massive altered uraninite, mobilizing part of its U, and precipitating the late colloform uraninite. This age is the same as that of the Antler orogenic event in the Canadian Cordillera (Burchfiel and Royden, 1991; Wright and Grove, 2008). The eastward thrusting during this orogenic event was associated with deformation and fluid circulation in the Western Canada Sedimentary Basin (Root, 2001), which extended over the Athabasca Basin (Fig. 1B). The Antler Orogeny may have also affected fluid-conducting fault zones in the Athabasca Basin, some of which were associated with the uranium deposits, because these are structurally hosted.

The paragenetically late uranophane gives two discordia lines with upper intercepts at ca. 460 and ca. 890 Ma (Fig. 7D), which could be interpreted as two stages of uranophane formation, and subsequent resetting at ca. 2 Ma (lower intercept ages), resulting in the loss of almost all radiogenic Pb. However, a formation age of ca. 460 or ca. 890 Ma contradicts the paragenetic position of the uranophane, which formed later than both the massive altered uraninite and the unaltered colloform uraninite (Fig. 2). Both uranophane and late coffinite form swarms of μm -wide veins that crosscut all other uranium minerals present. Therefore, the uranophane formation age is probably ca. 2 Ma, which is the same age as the lower intercept obtained on late colloform uraninite (Fig. 7D). This concurrence of ages clearly indicates that uranophane was formed during the same event (possibly Quaternary meteoric

fluids penetrating to the unconformity) that disturbed the isotopic system of the late colloform uraninite, causing partial loss of its radiogenic Pb. The few discordant analyses of uranophane (Fig. 7E) probably correspond to very limited amounts of radiogenic Pb incorporated into uranophane when it formed. The relatively large quantity of uranophane, not observed elsewhere in the Athabasca Basin, and the Devonian ages obtained on the massive uraninite underline the significant permeability that has existed at the Virgin River project. This permeability is probably related to the major Dufferin Lake Fault immediately to the west of the Centennial deposit (Fig. 1C).

Clay alteration fluids

As discussed above, the Centennial deposit was affected by several fluid circulation events. The main alteration- and U-related fluids in the Athabasca Basin are an evolved basinal brine and a reducing basement fluid, both of which are necessary to produce a deposit (e.g., Kotzer and Kyser, 1995). The character of the alteration fluids from stable isotopes analyses of the alteration clays could be interpreted as a mixing between two sources: (1) a high $\delta^{18}\text{O}$ and low δD fluid, recorded by kaolinite, that can either be low latitude, near coastal meteoric waters that have evolved as a result of extensive water-rock interaction, or a metamorphic or magmatic fluid (basement fluid), and (2) a lower $\delta^{18}\text{O}$ and higher δD fluid, recorded by chlorite and possibly corresponding to evolved seawater (Fig. 6). In other terms, the two fluids present are seawater and a fluid that either originated from the basement or acquired basement isotopic signature through extensive water-rock interaction. An alternative explanation could be the presence of a single fluid responsible for the pre-ore clay alterations, which would be a seawater modified by a varying degree of water-rock interaction: low water/rock ratio for kaolinite (acquiring "rock-dominated" signature), intermediate for illite, and high for chlorite (close to the original seawater signature; Fig. 6). Both explanations are plausible, but it should be noted that pre-ore alterations in the Athabasca Basin are produced predominantly by evolved seawater (e.g., Kotzer and Kyser, 1995).

Formation of the Centennial deposit

The diverse uranium mineral inventory, timing of the U mineralization-forming events, and fluid history of the Centennial deposit indicate that the deposit formed in a series of discrete stages. This is a departure from the evolution of unconformity-type U deposits in the Athabasca Basin, where remobilizations of U may occur locally, but are volumetrically insignificant. At Centennial, the main U stage, as represented by massive and variably altered uraninite, probably occurred as early as ca. 1.6 Ga, the same time as the disseminated uraninite observed at Virgin River-NNE (Figs. 2, 7). This U mineralization was followed by colloform uraninite at 384 ± 13 Ma and a late coffinite and uranophane stage at ca. 2.5 Ma. One of the causes for this recurrent U remobilization is most probably the high permeability of the sandstones in the Virgin River area, in part due to its proximity to the major Dufferin Lake Fault (Fig. 1C). The crosscutting of the orebody by the diabase dikes, rarely observed in the Athabasca Basin, might have further increased the permeability of the

immediate host rocks, leading to significant postdike fluid circulations.

Pb isotopes as an exploration tool

The utility of Pb isotopes in the exploration for uranium deposits in the Athabasca Basin has been demonstrated (Holk et al., 2003; Alexandre et al., 2009; Cloutier, 2009). The mobile and leachable Pb isotopes and U concentrations from the Virgin River area indicate that $^{207}\text{Pb}/^{206}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (Table 4) vary with proximity to a U deposit (Fig. 8). Even several km away, the Pb ratios in samples nearest the unconformity record a radiogenic source due to the strong permeability both of the basal portions of the Athabasca Group sediments and the highly permeable basement-sandstone unconformity (Cloutier, 2009). This is not observed in more distant parts of the Virgin River project, where samples from the lowermost sections of the Manitou Falls Formation taken only ca. 6 km to the north of the deposit (drill hole VR05, Fig. 1C) have common Pb signatures with $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of ca. 0.8 (Table 4). Samples from immediately above the unconformity and situated ca. 3 km to the east from the center of the Virgin River-NNE area (drill hole VR13, Fig. 1C) have radiogenic $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of ca. 0.6 (Table 4). The rapid increase in the $^{207}\text{Pb}/^{206}\text{Pb}$ ratios toward common Pb values of ca. 0.8 over a few km underlines the usefulness of Pb isotopes, effective even at the project scale.

Summary

The Centennial deposit, situated on the south-central margin of the Athabasca Basin, presents several characteristics of a typical unconformity-related deposit as defined in the basin (e.g., Jefferson et al., 2007), including mineralogy, chemical composition of the main alteration minerals, and origin of the fluids related to these alterations. However, the ore mineralogy of the deposit represents a clear departure from what has been observed elsewhere in the basin, in particular in the amounts of uranous and uranyl phases present at Centennial, reflecting a complex and protracted fluid history. The presence of the diabase dikes, probably part of the McKenzie igneous event, which crosscut the mineralization, not only gives an indication for the age of the main mineralization, situated at between 1.27 and 1.59 Ga, but also underlines the particular conditions favorable for the preservation of U ore here: in spite of a high permeability and the evidence of multiple fluid events, there is very little ore destruction observed at Centennial. This apparent contradiction between high permeability and multiple fluid circulation events on one side and uraninite ore preservation on the other can only be resolved by the character of the fluids involved, which can only have been reducing but also poor in complexing agents such as P, C, F, and Cl. Finally, the usefulness of Pb isotopes as an exploration tool is clearly demonstrated at Centennial, with low $^{207}\text{Pb}/^{206}\text{Pb}$ ratios, corresponding to radiogenic Pb, found in close proximity to the deposit.

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