

# UNCONFORMITY-ASSOCIATED URANIUM DEPOSITS OF THE ATHABASCA BASIN, SASKATCHEWAN AND ALBERTA

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## Abstract

This review of the geology, geophysics, and origin of the unconformity-associated uranium deposit type is focused on the Athabasca Basin. Pods, veins, and semimassive replacements of uraninite (var. pitchblende) are located close to unconformities between late Paleo- to Mesoproterozoic conglomeratic sandstone basins and metamorphosed basement rocks. The thin, overall flat-lying, and apparently unmetamorphosed but pervasively altered, mainly fluvial strata include red to pale tan quartzose conglomerate, sandstone, and mudstone. Beneath the basal unconformity, red hematitic and bleached clay-altered regolith grades down through chloritic altered to fresh basement gneiss. The highly metamorphosed interleaved Archean to Paleoproterozoic granitoid and supracrustal basement gneiss includes graphitic metapelitic that preferentially hosts reactivated shear zones and many deposits. A broad variety of deposit shapes, sizes and compositions ranges from monometallic and generally basement-hosted veins to polymetallic lenses located just above or straddling the unconformity, with variable Ni, Co, As, Pb and traces of Au, Pt, Cu, REEs, and Fe.

## Résumé

Cet examen de la géologie, de la géophysique et de l'origine des gîtes d'uranium associés à des discordances est focalisé sur le bassin d'Athabasca. Les minéralisations d'uraninite (variété pechblende) qui prennent la forme de lentilles, de filons ou de corps semi-massifs de remplacement se situent près de la discordance entre des grès conglomératiques de bassin du Paléoprotérozoïque tardif au Mésoprotérozoïque et les roches du socle métamorphosées. La succession sédimentaire de bassin est mince, repose dans l'ensemble à plat, est apparemment non métamorphosée, mais profondément altérée, et se compose principalement d'unités fluviales constituées de conglomérats quartzeux, de grès et de mudstones de couleur rouge à chamois pâle. Sous la discordance marquant la base de la succession sédimentaire, un régolite hématitique rouge, décoloré par endroits par une altération argileuse, passe progressivement vers les profondeurs du socle à des gneiss chloritisés puis à des gneiss non altérés. Les roches très métamorphosées du socle, formées d'une intercalation de gneiss granitoïdes et de gneiss supracrustaux de l'Archéen au Paléoprotérozoïque, incluent des métapelites graphitiques qui renferment de manière préférentielle des zones de cisaillement réactivées et un grand nombre de gîtes. Les gîtes sont de formes, de dimensions et de compositions très variées, passant de minéralisations monométalliques, généralement sous forme de filons encaissés dans le socle, à des lentilles polymétalliques présentant des concentrations variables de Ni, Co, As, Pb et des traces de Au, Pt, Cu, de terres rares et de Fe, qui se situent à cheval sur la discordance ou à peu de distance au-dessus.

## Introduction

This synopsis of unconformity-associated uranium (also unconformity-related and -type) deposits emphasizes the Athabasca Basin. The empirical term 'associated' is chosen because some genetic aspects are still under debate and the deposits occupy a wide range of spatial positions and shapes with respect to the unconformity. An expanded version of this paper introduces the final volume for EXTECH IV, Athabasca Uranium Multidisciplinary Study (Jefferson et al., 2007). Citations therefore include the most recent publications of EXTECH IV in addition to classic references.

After a concise definition, the grade, tonnage, and value statistics of unconformity-associated uranium deposits are provided in global and Canadian context. Geological attributes are summarized on continental, district, and deposit scales: favourable expressions of deposits; their size, morphology, and architecture; ore mineralogy and composition; and alteration mineralogy, geochemistry, and zonation. Key

exploration criteria are summarized for geology, geochemistry, and geophysics. Genetic and exploration models are reviewed in terms of conventional knowledge and recent advances, with reference to uranium sources, transport and focus of deposition. Conceptual and applied knowledge gaps are evaluated at the district and deposit scales. New lines of research and areas of uranium potential are proposed.

## Definition

Unconformity-associated uranium deposits are pods, veins, and semimassive replacements consisting of mainly uraninite dated mostly 1600 to 1350 Ma, and located close to basal unconformities between Proterozoic redbed basins and metamorphosed basement rocks, especially supracrustal gneiss with graphitic metapelite. Prospective basins in Canada (Figs. 1, 2) are 1 to 3 kilometres thick, relatively flat-lying, unmetamorphosed but pervasively altered, Proterozoic (ca. 1.8 to <1.55 Ga), mainly fluvial conglomeratic sandstone. The basement gneiss is paleoweathered with

variably preserved thicknesses of reddened, clay-altered hematitic regolith grading down through a green chloritic zone into fresh rock. Monometallic, generally basement-hosted ore pods, veins, and breccia in reactivated fault zones. Polymetallic, commonly subhorizontal ore lenses straddle the unconformity, replacing sandstone and altered basement rock with variable amounts of U, Ni, Co, and As, and traces of Au, PGEs, Cu, REEs, and Fe.

**Grade, Tonnage, and Value Statistics**

*Global Unconformity Resources*

World uranium resources are contained in some fourteen different types of deposits (Organization for Economic Co-operation and Development, Nuclear Energy Agency, and the International Atomic Energy, 2004), with Proterozoic unconformity deposits constituting more than 33%, mainly in Australia and Canada. Uranium resource data for Canadian and comparative Australian unconformity deposits are compiled with original references in the digital Canada Minerals Database by Gandhi (2007). Appendix 1 (CD-ROM) and Figure 3 summarize individual grades and tonnages of 42 Canadian and Australian deposits, illustrating their current relative importance. Table 1 provides totals for the Athabasca and Thelon basins. The Hornby Bay and Elu basins are less well explored and no unconformity resources have been outlined (Roscoe, 1984; Gandhi, 2007).

Unconformity deposits of the Athabasca Basin are the world's largest storehouse of high-grade U resources and are the sole producers of Canada's primary U. The most spectacular grades and tonnages (Appendix 1, Fig. 3) are those of Cigar Lake (east and west zones combined=875 kilotonnes of ore grading ~15% and containing 131,400 tonnes U) and McArthur River (1017 kilotonnes of ore grading ~22.28% and containing 192,085 tonnes U). The average

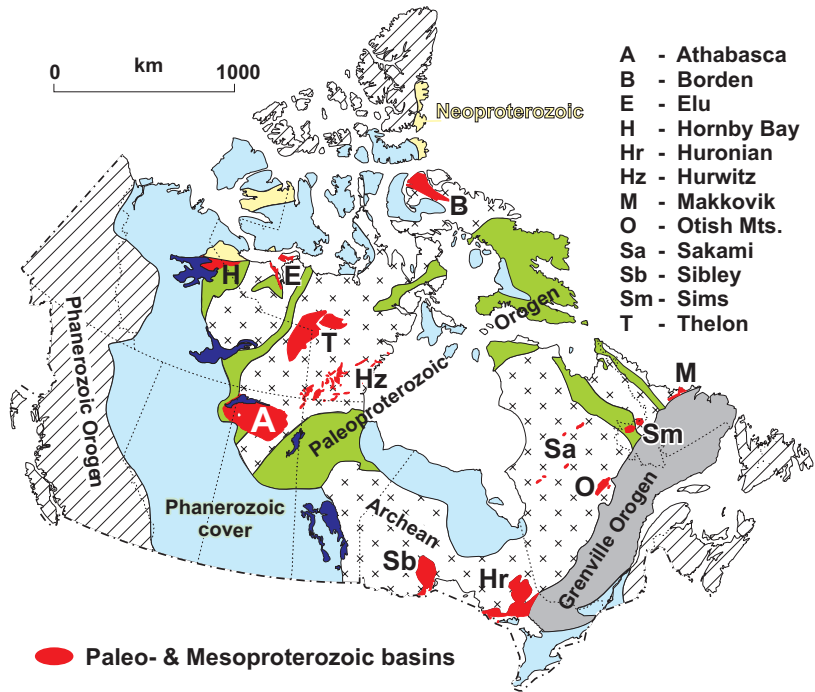


FIGURE 1. Paleo- to Mesoproterozoic basins within the Canadian Shield that contain unconformity-associated uranium deposits (e.g. Athabasca and Thelon) or are considered to have potential for them.

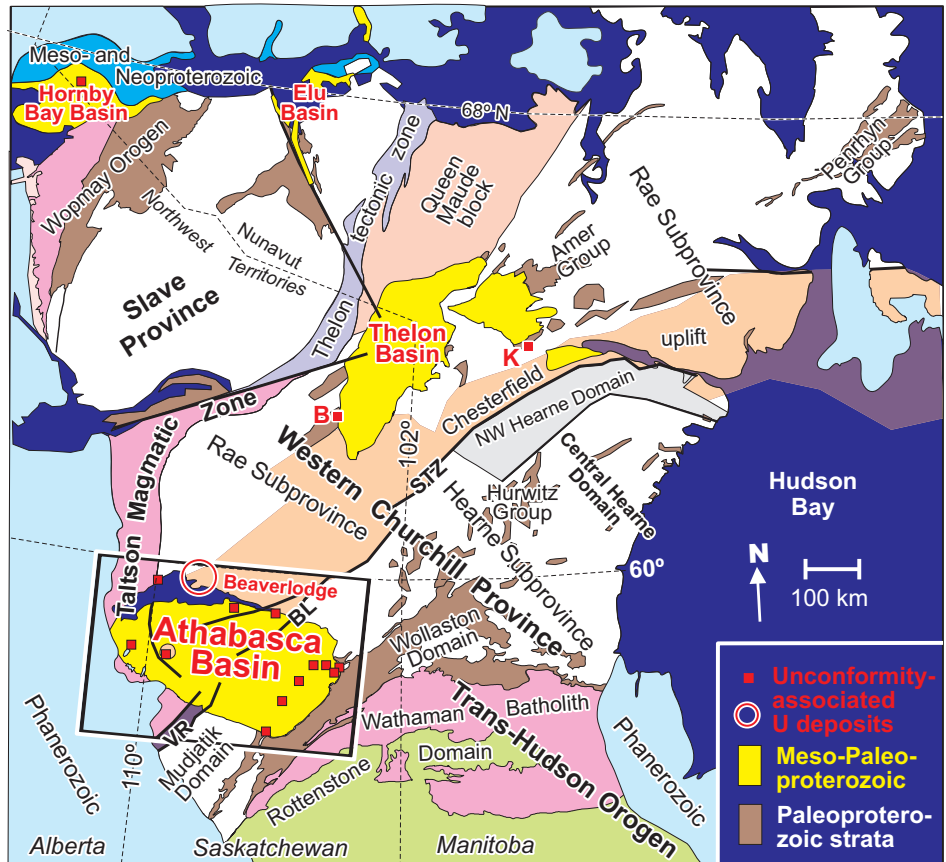
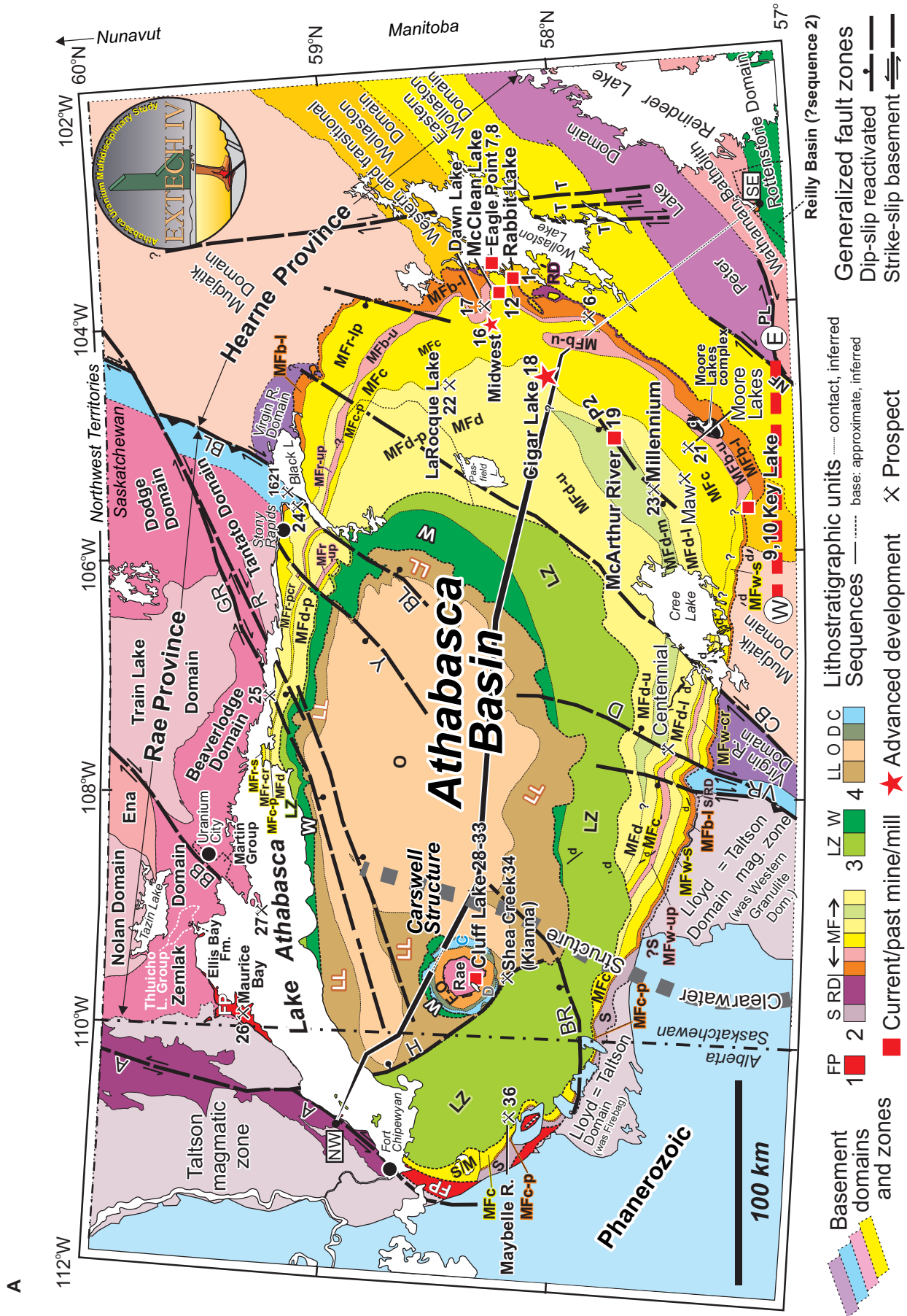
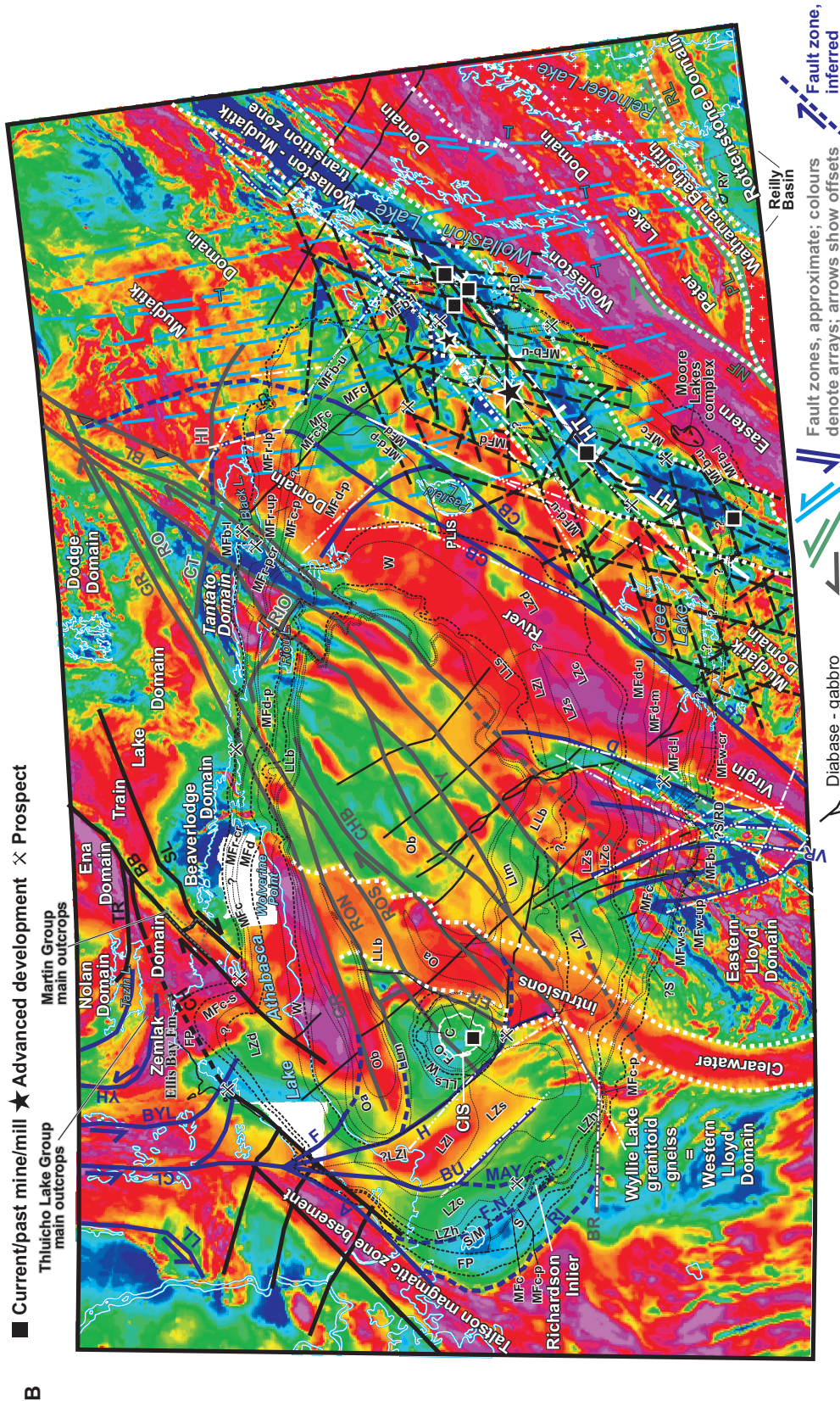


FIGURE 2. Location of unconformity-associated occurrences and the Athabasca Basin relative to major tectonic elements of the northwestern Canadian Shield, after Thomas et al. (2000) and Pehrsson (pers. comm., 2005). Occurrences are listed in the Appendix 1, known conventional resources are plotted in Figure 3, and Athabasca Basin occurrences are located in more detail on Figure 4A. Unconformity-associated prospects of the Thelon Basin are Boomerang Lake (B) and Kiggavik (K). Hornby Bay Basin (HBB) hosts the sandstone-hosted PEC-YUK prospect.







**FIGURE 4.** Geological setting and unconformity-associated uranium occurrences (numbered as in the Appendix on CD-ROM) of the Athabasca Basin region in northern Saskatchewan and Alberta, where Rae and Hearne are termed 'Province' rather than Subprovince (see Fig. 2). Symbols and fonts are slightly larger for more significant occurrences. Basement geology is after Portella and Annesley (2000a), Thomas et al. (2002), Card et al. (2003, 2007a, b), and Card (2006). Athabasca Basin geology and stratigraphic units (Table 3) are from Ramaekers et al. (2007). C=Carwell, D=Douglas, FP=Fair Point, LL=Locker Lake, LZ=Lazenby Lake, MF=Manitou Falls (members: b=Bird (l=lower, u=upper) e=Collins, d=Dunlop, t=Raibl (up=upper pebbly), w=Warmes (up=upper pebbly)), O=Otherside, RD=Read, S=Smart, W=Wolverine Point, d=diabase. Members of LZ, LL, and O are indicated by lines and labels but only one shade is used per formation. The western Wollaston Domain and the Wollaston - Mudjatik transition of Portella and Annesley (2000a, b) are combined here as "Wollaston-Mudjatik transition zone". CIS=Carwell Structure. Generalized fault zones after Ramaekers (2004) include multiple ductile movements before deposition of Athabasca Group and brittle transcurent and dip-slip movements during and after deposition; they are named as: A=Alan, BB=Black Bay, BL=Black Lake, BR=Beatty River, BU=Bustard, CB=Cable Bay, CH=Charlot, CHB=Charbonneau, CL=Charles Lake, CT=Clut, D=Dufferin, ER=East Rim, F=Fidler, FN=Fowler-Net Lake, GR=Grease River, H=Harrison, HT=Hudsonian thrusts (general trajectory), LL=Leland Lakes, MAY=Maybelle, NF=Needle Falls, PL=Parker Lake, P2=P2 fault at McArthur River, R=Richardson, RI=Riou, RL=Reilly Lake, RO=Robillard, RON=Robillard north, ROS=Robillard south, SL=St. Louis, T=Tabbemor, VR=Virgin River array (Dufferin is one named fault of many in VR), Y=Yaworski, YH=Yaisore-Hill Island. Arrays of faults with similar orientation and offset are indicated by colour groups. A) Simplified bedrock geology. Cross-sections of Figure 5A and B are located along dotted lines labelled NW-SE (along the basin axis) and E-W (south of Key Lake). B) Outlines of stratigraphic units of the Athabasca Group (black), basement domains (white outlines) and major reactivated fault systems (heavy coloured lines) on total magnetic field (Geological Survey of Canada, 1987; Pilkington, 1989). Faults are after Portella and Annesley (2000a), Ramaekers (2004), Card et al. (2007a), and Thomas and McHardy (2007). Many late faults have limited offsets that cannot be shown at this scale (see Card, 2006).

**TABLE 1.** Summary of uranium resources in major Paleo- and Mesoproterozoic districts of northwestern Canada (shaded) and Australia; data from Appendix 1.

| District                          | Kt Ore <sup>1</sup> | % U <sup>2</sup> | Tonnes U |
|-----------------------------------|---------------------|------------------|----------|
| Athabasca Basin                   | 29,811              | 1.97             | 587,063  |
| Beaverlodge District <sup>3</sup> | 15,717              | 0.165            | 25,939   |
| Thelon Basin                      | 11,989              | 0.405            | 48,510   |
| Hornby Bay Basin                  | 900                 | 0.3              | 2,700    |
| Kombolgie Basin                   | 87,815              | 0.323            | 283,304  |
| Paterson Terrane                  | 12,200              | 0.25             | 30.5     |
| Olympic Dam <sup>4</sup>          | 2,877,610           | 0.03             | 863,283  |

1. Includes past production.
2. Calculated from Kt ore and tonnes uranium, rounded to significant digits.
3. Past production from two “classic vein-type” (Eldorado and Lorado Mills) and one episyenite-type (Gunnar) deposits.
4. Genetically linked with the 1850 Ma Gawler Range volcano-plutonic complex. Olympic Dam is breccia hosted, not unconformity-associated, but is included here for comparison because it is such a vast individual resource of uranium, of approximately the same age as the unconformity-associated deposits listed here (references in Gandhi, 2007).

Uranium ore veins are present only in the basement near its basal unconformity and are absent where the Martin Group pinches out, despite continuity of favourable basement lithologic units and structures (e.g. Robinson, 1955, p. 73-74; Tremblay, 1968; Mazimhaka and Hendry, 1989; T. Trueman, oral presentation in Saskatoon, November 2005). Exploration is now reevaluating this district.

*Canadian Unconformity-Associated Uranium Resources – Potential Producers*

The Thelon Basin is very similar in size, general age, and geological attributes to the Athabasca Basin (Figs. 1, 2; Table 2), yet its reasonably assured U resources are 9% of the Athabasca Basin. These are concentrated in basement-hosted deposits of the Kiggavik trend, an area less than 500 km<sup>2</sup> with an average grade similar to that of the Kombolgie Basin. Exploration is active throughout the Thelon Basin, except in the Thelon Game Sanctuary.

The term ‘Hornby Bay Basin’ refers to the region outlined by exposures of the Hornby Bay Group. This restricted usage excludes the ‘Amundsen Basin’ (regional stratigraphy defined by Rainbird et al., 1994) that, although used by Kerans et al. (1981) as including Hornby Bay Group, is

too broad to provide metallogenic discrimination (see below). The only deposit in Hornby Bay Basin with measured resources (PEC prospect, Appendix 1, Fig. 3) is classified as sandstone type (Bell, 1996) and is the only significant example of this deposit type in the Proterozoic of Canada. Nonetheless, the Hornby Bay Basin is viewed as correlative with the Athabasca and Thelon basins, and is being intensively explored for unconformity deposits. The Elu Basin, extending from Bathurst Inlet north to Hadley Bay on Victoria Island (Figs. 1, 2), is broadly correlative with the Hornby Bay Basin (Campbell, 1979). Additional prospective basins, such as Huronian, upper Hurwitz, Otish, Sibley, and Sims, are discussed under *Ages of Known and Prospective Districts* below.

**Geological Attributes**

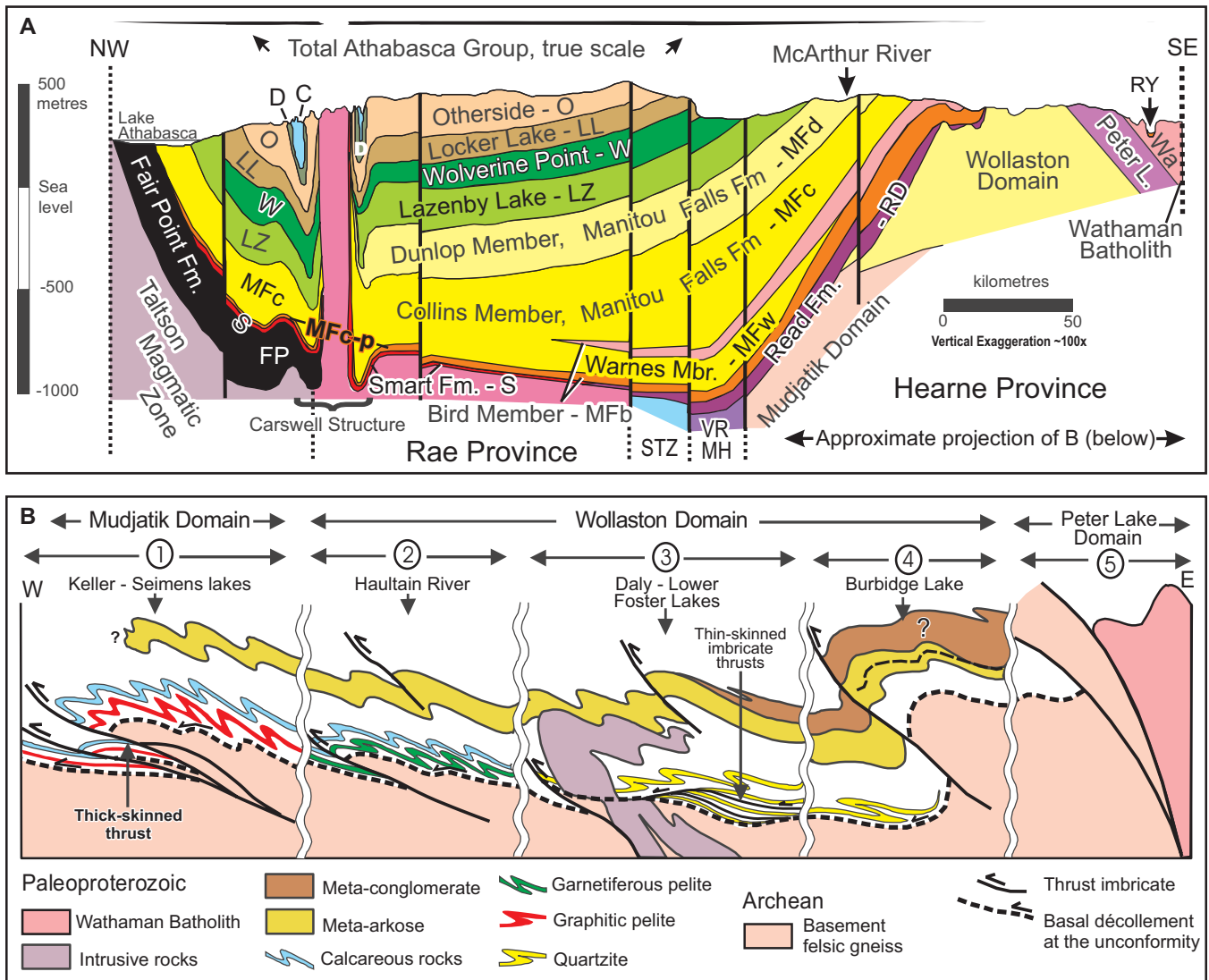
*Continental-Scale Geological Attributes*

The continental-scale geotectonic environment of significant unconformity-associated U deposits is at the base of flat-lying, fluvial redbed strata on peneplaned tectonometamorphic complexes in the interiors of large cratons. The Athabasca and Thelon basins are located on the western Churchill Province between the eroded remnants of two major orogenic belts: the ca. 1.9 Ga Taltson magmatic zone to Thelon tectonic zone and the ca. 1.8 Ga Trans-Hudson Orogen (Fig. 2). These belts accommodated ductile transpression during convergence of the Slave and Superior provinces (e.g. Hoffman, 1988) and form the western and eastern portions of the Rae and Hearne provinces, respectively. High heat flow could have resulted from the voluminous radiogenic granitoid intrusions within these orogens.

**TABLE 2.** Comparison of Athabasca and Thelon basins (after Miller and LeCheminant, 1985; Gandhi, 1989; Kyser et al., 2000).

| Attribute  | Athabasca        | Thelon          |
|--|------------------|-----------------|
| Graphitic metasedimentary rocks beneath ore      | Distinct         | Locally         |
| Paleoweathering profile below basal unconformity | Shallow to deep  | Shallow to deep |
| Subbasins developed via reactivated faults       | Yes              | Yes             |
| Maximum age of sedimentation (Ma)                | ca. 1720-1750    | ca.1720         |
| Fluorapatite                                     | Yes              | Yes             |
| Aeolian sandstone                                | Possible         | Yes             |
| Arkosic sandstone regionally clay altered        | Minor            | Yes             |
| Quartz overgrowths preserve hematite rims        | Yes              | Yes             |
| Early detrital kaolin in matrix                  | Yes              | No?             |
| Peak diagenetic clay minerals                    | Dickite + illite | Illite          |
| Peak diagenetic / hydrothermal temperatures      | ~240°            | ~200°           |
| Illite incorporates Mg and Fe                    | in regolith only | Variable        |
| Corroded zircon grains near ore zones            | Local            | No?             |
| Regional fresh zircon                            | Yes              | Yes             |
| Extensive aluminum phosphate ± sulphate          | Yes              | Yes             |
| potassium-feldspar + chlorite at 1 Ga            | No               | Yes             |
| Late vein carbonate from meteoric water          | Yes              | Yes             |
| Bleaching and clay alteration halos              | Yes              | Yes             |
| Sandstone / unconformity-hosted uranium          | Yes              | One example     |
| Basement-hosted uranium                          | Yes              | Yes             |
| Significant deposits                             | Yes              | One             |

Unconformity-Associated Uranium Deposits of the Athabasca Basin, Saskatchewan and Alberta



**FIGURE 5. A)** Lithostratigraphic cross-section of the Athabasca Basin, after Ramaekers (1990) and Ramaekers et al. (2007). True 1:1 scale is shown at top. Stratigraphic units in the vertically exaggerated section are formations except those starting with MF, which are members of Manitou Falls Formation; MFw is Warnes Member. Basement domains are diagrammatic. Line of section is shown as NW-SE in Figure 4A. **B)** Diagrammatic structural cross-section south of Key Lake, adapted from Tran (2001) and Györfi et al. (2007), illustrates structural geometry of the Wollaston-Mudjatik transition zone that underlies the most economically productive area of the eastern Athabasca Basin. Line of section is shown as W-E in Figure 4A. The circled numbers 1 through 5 refer to discrete areas mapped by Tran (2001) from which the components of this diagram were obtained.

There is no suggestion that the Athabasca Basin region was a site of enhanced mafic magmatism (Buchan and Ernst, 2004).

Original thicknesses and lateral extents of the most prospective basins were somewhat greater than what is preserved, but far less than foreland or continent-margin basins. The Canadian basins are interpreted as discrete ‘lakes of gravel and sand’ separated by large areas of limited accommodation (Ramaekers et al., 2007). The maximum cored thickness of the Athabasca Group in any one place is about 1500 m – the four major depositional sequences separated by basin-wide unconformities (Table 3) record repeated deposition and erosion during about 200 Ma. Shallow marine strata are minor or cap the redbed sequences (Figs. 4, 5).

The Thelon Basin developed in an interior position, very much like the Athabasca Basin (Rainbird et al., 2003a), with some 1800 m (Overton, 1977) comprising mainly the Thelon Formation with three depositional fluvial sequences (Hiatt et

al., 2003) capped by the thin, volcanic Kuungmi and carbonate Lookout Point formations (Cecile, 1973; Rainbird et al. 2003a). In the Hornby Basin, paleocurrents from the west for basal units (Kerans et al., 1981) indicate the Hornby Bay Basin also formed within an intracratonic setting, but thicknesses are greater and more variable, with evidence for syn-depositional compression (MacLean and Cook, 2004).

The older, past-producing Martin Basin north of Lake Athabasca (Mazimhaka and Hendry, 1989) and the prospective Baker Lake Basin east of Thelon Basin (Miller, 1980) are distinct pull-apart structures filled by relatively thick but laterally restricted sequences and host smaller, lower grade, unconformity-associated deposits and occurrences. The younger, Amundsen Embayment, located northwest of Hornby Bay and Elu basins (Young et al., 1979), had a marginal marine setting with no coarse siliciclastic rocks, layercake stratigraphy, and a metallogeny dominated by sedimen-

**TABLE 3.** Lithostratigraphic units and unconformity-bounded sequences of the Athabasca Group (after Ramaekers et al., 2007). The framework mineral for all textural types from conglomerate to mudstone is 99% quartz. Every unit contains crossbedding and ripple crosslamination, and most contain single-layer thick quartz pebble or granule beds. Only diagnostic stratigraphic parameters, such as grain-size and desiccation cracks, are summarized here. Aggregate maximum thickness of each formation is summarized in metres in the left-hand column. Inferred minimum age of basement and U-Pb age of volcanic zircon from Wolverine Point are by Rainbird et al. (2007). Douglas Formation Re-Os age is by Creaser and Stasiuk (2007). <sup>S</sup> = present only in Saskatchewan.

| Formation<br>[code]  | Member; [code] (textural lithology)<br>MTG = maximum transported grain size in mm   | Sequence     |              |
|--|---|--------------|--------------|
| Carswell<br>[C] 500 m  | Upper and lower carbonate (dololutite, dolorudite, stromatolite, oolite, dolarenite) <sup>S</sup>   | 4            |              |
| Douglas<br>[D] 300 m   | (dark grey carbonaceous mudstone with desiccation or sphaerulite cracks and interbeds of fine to very fine quartzarenite, MTG <2). Organic matter is 1541 ± 13 Ma by Re-Os isochron <sup>S</sup>  |              |              |
| Now recognized as up-faulted Locker Lake Formation <sup>S</sup>  |   |              |              |
| Otherside<br>[O]<br>183 m  | Birkbeck [Ob] (quartzarenite with minor thin interbeds of dark mudstone near the top; MTG <2 except for pebbly unit near base)<br>Archibald [Oa] (quartz-pebbly quartzarenite, quartzarenite; MTG<8)  |              |              |
| Locker<br>Lake [LL]<br>288 m   | Marsin [LLm] (quartz-pebbly quartzarenite; MTG 8-16)<br>Brudell [LLb] (thin conglomerate beds in quartzarenite; MTG >16)<br>Snare [LLs] (quartz-pebbly quartzarenite; MTG 2-16, sparse mudstone interbeds <50 cm thick)   |              |              |
| Unconformity   |   |              | Unconformity |
| Wolverine<br>Point [W]<br>186 m  | Claussen [Wc] (interstitial-clay-rich quartzarenite, sparse mudstone interbeds <1 m thick; MTG <2)<br>Brule [Wb] (interbedded mudstone >50 cm and tuffaceous quartzarenite, common thin intraclast conglomerate; MTG <2 except for local basal lag. Zircon in tuff intraclasts is 1644 ± 13 Ma by U-Pb.   |              | 3            |
| Lazenby<br>Lake [LZ]<br>460 m<br>(aggregate<br>thickness<br>excludes<br>laterally<br>equivalent<br>Dowler<br>Member)                 | Dowler [LZd] (quartzarenite, minor siltstone and quartz-pebbly quartzarenite; MTG < 8)<br>Larter [LZl] (quartz-pebbly quartzarenite, minor mudstone intraclasts; MTG <8)<br>Shiels [LZs] (quartz-pebbly quartzarenite with pebbly layers, rare mudstone beds and intraclasts; MTG >8)<br>Clampitt [LZc] (pebbly base, quartz-pebbly quartzarenite, minor laminated siltstone and mudstone; MTG <8)<br>Hodge [LZh] (5-30 cm basal conglomerate, quartz-pebbly quartzarenite and conglomerate, sandstone intraclasts; MTG >8)   |              |              |
| Basal unconformity to Mirror Basin   |   | Unconformity |              |
| Manitou<br>Falls [MF]<br>991 m<br>(thickness<br>excludes<br>Warnes<br>and Raibl<br>members<br>laterally<br>equivalent<br>to Bird)    | Dunlop [MFd] (>1% clay-intraclasts in quartzarenite, mudstone interbeds; MTG <2) <sup>S</sup><br>Collins [MFc] (quartzarenite with minor quartz pebbly beds, mudstone interbeds, <1% clay intraclasts, <2% conglomerate interbeds)<br>Warnes [MFw] (quartzarenite and clay-intraclast-rich quartzarenite in Karras Deposystem, from Virgin River area to Alberta)<br>Raibl [MFr] (pebbly quartzarenite in Moosonees Deposystem, northeastern Athabasca Basin; minor clay intraclasts, <2% conglomerate; MTG >2) <sup>S</sup><br>Bird [MFB] (interbedded >2% quartz-pebble conglomerate, quartz-pebbly quartzarenite, thin mudstone and siltstone interbeds; MTG >2) | 2            |              |
| F-O: Undivided Fair Point to Otherside formations in Carswell Structure <sup>S</sup>   |   |              |              |
| Local unconformity separates Manitou Falls and Read formations <sup>S</sup>  |   |              |              |
| S/M: undivided Smart or Manitou Falls formations (only in Alberta)   |   |              |              |
| Smart [S]<br>153 m   | (quartzarenite with local red mudstone and oncolite interbeds at base). May be a distal equivalent of Read Formation  | 1            |              |
| Read [RD]<br>156 m   | (basal quartz>lithic pebble conglomerate, interbedded low-angle bedded quartzarenite, quartz-pebbly quartzarenite and quartz pebble conglomerate, common but local red quartz siltstone to mudstone intraclasts and interbeds with desiccation cracks; MTG >2) <sup>S</sup>   |              |              |
| Reilly<br>[RY]   | RYcg (conglomeratic quartzarenite) <sup>S</sup>   |              |              |
| Basal unconformity to Reilly Basin <sup>S</sup>  |   | Unconformity |              |
| Fair Point<br>[FP]<br>>380 m   | Beartooth [FPb] (0-10% quartz>lithic-pebbly quartzarenite with abundant matrix clay; MTG generally <64)<br>Lobstick [FPl] (interbedded >2% quartz>lithic conglomerate, quartz>lithic-pebbly quartzarenite and local basal quartz-pebbly red mudstone with minor desiccation cracks; MTG commonly >64)   | 1            |              |
| No formal designation  | BL (basal lag, pebbles to boulders)   | Unconformity |              |
| Basement: interleaved Archean granitoid gneiss, Paleoproterozoic paragneiss, late intrusions and metamorphism (titanite ca. 1750 Ma) |   |              |              |



tary and volcanic copper, rather than U (Delaney et al., 1982).

The influences of plate or plume tectonics on the origins, diagenesis, and mineralization of intracontinental Proterozoic basins have been considered enigmatic (Ross, 2000). Peneplaned and deeply paleoweathered basement gneiss and the relatively thin cover recorded by flat-lying basins of continental strata imply a relatively stable tectonic environment that persisted from before to long after ore deposition. Despite the internal continental setting, complexities within these basins tell a story of subtle but highly influential transpressional and extensional tectonics that reactivated faults vital to ore formation (see the following section and Ramaekers (2004)).

### *District-Scale Geological Attributes*

#### Unconformities

The first-order favourable attribute is the unconformity at the base of a relatively flat-lying and intracontinental, unmetamorphosed, late Paleoproterozoic to Mesoproterozoic, fluvial, conglomeratic sandstone, diagenetic redbed sequence. Prospective Proterozoic basins in Canada and Australia are typically underlain by extensive red hematitic paleoregoliths (Fraser et al., 1970; Cecile, 1973; Macdonald, 1980, 1985; Miller et al., 1989; Ramaekers, 1990; Gall, 1994). Variable thicknesses of regolith grade down through green chloritic altered rock into fresh basement (Hoeve and Qurt, 1984; Kyser et al., 2000). The regolith is interpreted as a result of regional paleoweathering (see review by Gall and Donaldson, 2006) that has been overprinted by diagenetic bleaching and additional hematite alteration (Macdonald, 1985).

The basement immediately below the Athabasca Group has a vertical paleoweathered profile ranging from a few centimetres up to 220 metres thick, with some deeper pockets and slivers developed along fault zones (Macdonald, 1980). Regionally, the upper portion of the regolith profile exhibits strong red hematitic alteration that grades downward through greenish chloritic alteration into fresh basement rock. Extensively within mineralized districts, a bleached zone variably overprints the top of the red zone of the paleoweathered basement and basal units of the Athabasca Group. The bleached zone comprises buff-coloured clay and quartz. It crosscuts and therefore postdates the red zone beneath. In profiles developed on the basement meta-arkose, a zone of white clay replacement of feldspar and mafic minerals separates the red and green zones. A downward progression from kaolinite to illite and chlorite is common through the regolith profile. Earthy bright red to nearly black crystalline hematite alteration in turn overprints the white alteration in mineralized areas.

In the western part of the Athabasca Basin, the unconformity between the Fair Point Formation and overlying Smart Formation is also marked by local red hematitic alteration (Ramaekers et al., 2007). Therefore, the regolith beneath the eastern part of the Athabasca Basin would have resulted from paleoweathering lasting from before deposition of the Fair Point Formation to before deposition of the Read and Manitou Falls formations, although much of the earlier paleoweathered material was probably transported into the

Jackfish Basin as a component of the Fair Point detritus. This is consistent with Macdonald's (1980) suggestion that the upper, soil-textured portion of the paleosol was mostly eroded before deposition of the Athabasca Group. Alternative views and additional details are presented by Jefferson et al. (2007).

#### Uranium-Rich Basement Complexes

Many workers have emphasized the favourability of basement domains that have U-enriched granite and pegmatite that were generated during regional high-grade metamorphism and anatexis of metasedimentary rocks (e.g. Thomas, 1983, Annesley et al., 1997; Madore et al., 2000; Cuney et al., 2003; Freiburger and Cuney, 2003; Hecht and Cuney, 2003). Abundant U-bearing minerals in such domains include monazite, zircon, and uraninite. The latter also forms numerous small prospects in pegmatite (Thomas, 1983) that are recorded in the Saskatchewan mineral database, west of the Needle Falls shear zone (Saskatchewan Industry and Resources, 2005). The radiogenic domains, also rich in K, Th, and rare earth elements (REE), may have selectively contributed U to the overlying basins through a variety of mechanisms (see *Genetic Models* below).

In the eastern part of the Athabasca Basin, the great majority of mines and prospects are located where the Athabasca Group unconformably overlies the transition between the western Wollaston and eastern Mudjatik basement domains (Fig. 2; Thomas, 1983; Annesley et al., 2005). This transition contains high proportions of pelitic, quartzose, and arkosic paragneiss that are isoclinally folded and interleaved with Archean orthogneiss and intruded by abundant pegmatite. Many significant deposits in this eastern area are also located at the metamorphosed unconformable contact between the Archean granitoid gneiss and late Paleoproterozoic basal Wollaston Supergroup (Yeo and Delaney, 2007), where it contains graphitic metapelitic gneiss (Annesley et al., 2005). The metapelitic constitutes a weak zone that focused deformation, including a major D1 décollement (Fig. 5B), during folding and thrusting (Tran, 2001).

Significant but fewer mined deposits and prospects are also located in the basement complex of the Cluff Lake area that was exposed by the central uplift of the Carswell Structure (Lainé et al. 1985), most again associated with graphitic units and close to the overturned basal unconformity of the Athabasca Group. High-grade intersections have been reported from other western localities in the Athabasca Basin, also associated with graphitic shear zones in supracrustal belts in the underlying basement. These are summarized by Rippert et al. (2000), Brouand et al., (2003), Card (2006), Wheatley and Cutts, (2006), Card et al. (2007), Kupsch and Catuneanu (2007), and Pan et al. (2007).

#### Reactivated Faults

The close relationship between unconformity-associated U deposits and faults has been known since the classic reports on the Rabbit Lake deposit by Hoeve and Sibbald (1978) and Hoeve et al. (1980), and are targets of exploration in all Canadian basins. Faults in the Athabasca Basin constitute a number of arrays (Figs. 4, 5) with different attributes, such as dextral or sinistral, extensional or transpressional,

and ductile or brittle, within which subsidiary splays may be invisible at the district scale but critical at the deposit scale. A number of originally ductile faults underwent repeated brittle reactivation, with offset on the order of tens to hundreds of metres, and were important for focusing mineralizing fluids. The largest of the reactivated faults offset the unconformity by hundreds of metres, e.g., the Dufferin Fault, whereas primary ductile offsets of basement units were at least tens of kilometres. For example, the P2 reverse fault offsets the unconformity by 20 to 40 m and at times was extensional (Bernier, 2004). Regional (Hajnal et al., 2007) to detailed (Györfi et al., 2007) analysis of seismic data shows the deep listric nature of the P2 fault zone and its geometric relationship to folds and thrusts of the Hudsonian Orogeny. Regional aeromagnetic data (Portella and Annesley 2000a, b; Ramaekers et al., 2007) show that many other prospective structures affect the Athabasca Basin.

These reactivated fault arrays are spatially linked with thickness, facies, and paleocurrent changes in the Athabasca Group (Ramaekers et al., 2007; Yeo et al., 2007). The fault zones served as hinge lines during some 200 million years of alternating sedimentation and erosion. Changes in basin polarity and fault valve activity would have influenced basin fluid flow, possibly up-dip toward basin margins (Hiatt and Kyser, 2007), along paleochannel conglomerate units (Collier and Yeo, 2001; Long, 2007), but mainly vertically along the fault conduits causing basement alteration and mineralization (e.g. Hoeve and Quirt, 1984).

#### District-Scale Graphitic Metapelite Gneiss

Graphitic basement units are key empirical exploration parameters for U deposits in the Athabasca Basin, and underlie the northeastern and southwestern Thelon Basin (e.g. Boomerang Lake, Fig. 2) (Davidson and Gandhi, 1989) and Kombolgie Basin, Australia (Dahlkamp, 1993). Graphitic metapelite is absent beneath the eastern part of the Thelon Basin along the Kiggavik Trend (Fig. 2) although the basement-hosted deposits are located mainly in supracrustal rocks (Fuchs and Hilger, 1989). Underlying the eastern Athabasca Basin, graphitic units are stratigraphically low in the metasedimentary components of the western Wollaston and eastern Mudjatik basement domains (Fig. 5B); these are members of the Karin Lake Formation, Daly Lake Group, Wollaston Supergroup (Yeo and Delaney, 2007). The protolith of the graphitic metapelitic gneiss appears to unconformably overlie older granitoid gneiss and forms the basal interface of the overlying Wollaston Supergroup. In a tectonostratigraphic reconstruction by Tran (2001), it is interpreted to be part of an east-west facies change in basal Wollaston Supergroup units from metaquartzitic gneiss, through garnetiferous silicate-facies iron formation to weakly sulphidic graphitic metapelite (e.g. Fig. 5B). Graphitic units underlying the western Thelon Basin are identified as part of the Amer Group (Miller and LeCheminant, 1985).

Graphitic metapelitic gneiss units in the Wollaston and Mudjatik domains constitute weak zones between competent units and were foci for local deformation during regional folding, thrusting, and later brittle deformation. Similar graphitic units underlie deposits in the Maybelle River (Pană et al., 2007) and Shea Creek areas of the western Athabasca

Basin where they are detectable as conductors and considered as components of supracrustal belts in the Taltson Magmatic Zone (Rippert et al., 2000; Brouand et al., 2003; Card et al., 2007a). These supracrustal components can be traced further east and northeast under the Athabasca Basin toward the Virgin River shear zone (Stern et al., 2003; Card, 2006; Card et al., 2007b). In the Cluff Lake area of the Carswell Structure, the graphitic metapelite and host metasedimentary units were interpreted as part of the Rae Province, such as those exposed north of the Athabasca Basin (Harper, 1982). Recent geochronology shows that magmatism coeval with the Taltson magmatic zone also affected both of these regions (S. Pehrsson, pers. comm., 2006). The graphitic metapelite units in these belts perhaps conducted deep crustal heat upward to the base of the Athabasca Basin where their thermal anomalies drove convection of ore-forming hydrothermal fluids (Hoeve and Quirt, 1984).

#### Quartz-Dominated, Uranium-Depleted Strata

The conglomeratic sandstone bodies that overlie the regolith and host parts of the polymetallic U deposits are thoroughly oxidized terrestrial redbed sequences with very long and complex diagenetic/hydrothermal-alteration histories (Table 2, Fig. 6). The preserved detrital framework in the Athabasca Basin is more than 99% quartz but much of the sand-grade material has only moderate textural maturity (Hoeve and Quirt, 1984; Ramaekers, 1990; Bernier, 2004; and Collier, 2004). Such compositional maturity in the absence of textural maturity, particularly in conglomeratic units such as the Fair Point, Read, and lower Manitou Falls formations, is typical of primary quartz-dominated detrital mineralogy that developed under a warm, tropical climate and did not require second- or third-cycle reworking (Dal Cin, 1968; Long, 2007).

Whatever the primary depositional mineralogy, the approximately 200 Ma duration of high temperature diagenesis in the Athabasca Basin ensured that any primary feldspar (Sibbald et al. 1976; Quirt, 1985) was altered to clay. Hiatt and Kyser (2000) summarized well established examples showing destruction of feldspar and other unstable minerals during burial and later diagenesis (Millikan, 1988; McBride, 1989). Nevertheless, in the bulk of the Athabasca Group, clay minerals form less than 3% of the rock, therefore if there was primary feldspar it was a very minor component (Macdonald, 1980; Kyser et al., 2000; and Wasyluk, 2002).

In contrast, variable primary feldspar and lithic fragments are preserved in quartz- and phosphate-cemented units of the Thelon Formation (Kyser et al., 2000) whose coarser grained units have been described as lithic arkose and subarkose (Cecile, 1973; Miller, 1983). Much of the Hornby Bay Group is feldspathic (Kerans et al., 1981). Nevertheless, the sediments filling these basins are also depleted in U as indicated by airborne gamma-ray surveys.

Evidence of preexisting unstable detrital minerals in the Athabasca basin is discussed in detail by Jefferson et al. (2007). Particular attention is paid to altered mafic heavy mineral laminae that now consist of zircon and quartz framework grains surrounded by an abundant irregular matrix of secondary hematite, clay minerals, and Th-rich aluminum

phosphate ± sulphate (AP) minerals, one being identified as florencite (Percival, 1989; Mwenifumbo and Bernius, 2007). These AP minerals are regionally abundant and relatively radioactive due to their Th content but essentially lack U.

District-Scale Alteration

In addition to paleoweathering and hydrothermal alteration of basement gneiss below the unconformity, two types of regional-scale alteration have been recognized: basin-wide pre-ore diagenetic sandstone alteration, and subbasin-scale alteration halos that outline trends and clusters of U deposits. One of the earliest recognizable regional diagenetic events in the Athabasca Basin is a pre-ore quartz overgrowth (Q1 event) that encapsulates hematite-coated detrital quartz grains. Extensive pressure solution is recorded by abundant stylolites, particularly in the Fair Point Formation (Sequence 1).

The Q1 event was followed by a complex diagenetic sequence that differs in the Athabasca, Thelon, and Kombolgie basins (Fig. 6, after Kyser et al., 2000), but is independent of stratigraphy. The original mixed clay matrix of the Athabasca Basin was shown by Hoeve and Quirt (1984) to have been dominantly kaolinitic with small amounts of montmorillonite, a range of chlorite minerals, and a low-Mg-Fe illite (after Nickel and Nichols, 1991). Regional diagenesis converted this to a mixture dominated by dickite, a higher crystallinity polymorph of kaolinite, with minor amounts of illite and chlorite (Quirt and Wasyluk, 1997; Earle et al., 1999; Quirt, 2001; Wasyluk, 2002). Hoeve and Quirt (1984) used the illite crystallinity to confirm the 200°C burial diagenetic temperature determined from fluid inclusions (Pagel, 1975), as well as to show increased diagenesis with depth in the Rumpel Lake borehole that cores the lower Locker Lake and complete Wolverine Point, Lazenby Lake, and Manitou Falls formations in the east-central part of the basin. The dominant clay mineral analysed by Kyser et al. (2000) and Renac et al. (2002) in the Thelon and Kombolgie basins is illite. Applying the calculations of Hoeve and Quirt (1984) to these basins suggests that sufficient K was present in detrital feldspar and mica to support complete conversion of detrital kaolinite to illite during their diagenesis.

A variation in the regional background dickite of the Athabasca Basin was noted by Earle and Sopuck (1989) in its southeastern part where a large illite anomaly forms a corridor, 10 to 20 km wide, that extends for 100 km northeast from Key Lake to Cigar Lake (Fig. 7). Earle et al. (1999) described the illitic alteration at Key Lake in more detail. The axis of this regional illite anomaly also contains subpar-

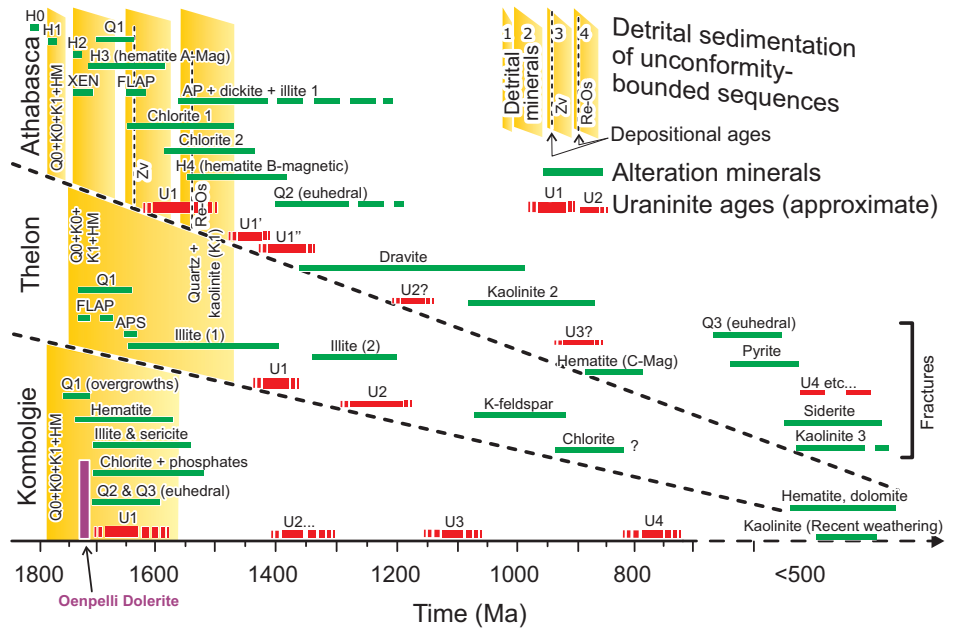


FIGURE 6. Simplified mineral paragenesis of the Paleo- to Mesoproterozoic Athabasca, Thelon, and Kombolgie basins, modified after Kyser et al. (2000, their Fig. 10.21) and Polito et al. (2004, 2005). The ages of ‘primary’ uraninite (the first deposition of uranium in each deposit) may be different for different deposits, hence U1 (ca. 1500-1600 Ma) and U1’-U1’’ ages (e.g. ca. 1350-1400 Ma) might both be primary in the Athabasca Basin; younger ages may be a result of remobilization and addition of uranium over time. Depositional ages in the Athabasca Group are Zv: U-Pb on volcanic zircon in Wolverine Point Formation (Rainbird et al., 2007), and Os-Re: primary organic matter in Douglas Formation (Creaser and Stasiuk, 2007). The ages of depositional sequences 1 through 4 for the Athabasca Group and of uranium deposits are summarized from discussions in text. The  $1723 \pm 6$  Ma age of the Oenpelli Dolerite is by  $^{40}\text{Ar}/^{39}\text{Ar}$  (Kyser et al., 2000) and U-Pb (Edgecombe et al., 2002). AP=aluminum phosphate; APS=aluminum sulphate phosphate; FLAP=fluorapatite; H0 is primary hematite in the paleoweathered regolith; H1 and H2 are very early diagenetic hematite in basal red mudstone beds; XEN=xenotime.

allel linear zones of anomalous chlorite and dravite. The dravite is clay sized, blue-green, and concentrated along fractures and disseminated in altered zones, overprinting illite, chlorite, and kaolinite, recording late hydrothermal boron metasomatism. The illite anomaly encompasses all known U deposits and prospects in the southeastern part of the basin (Fig. 4A), notably Key Lake, P-Patch (4 km east of Key Lake), McArthur River, BJ (7.5 km southeast of McArthur River), the Millennium prospect, and is discontinuous around the Cigar Lake mine and the Dawn Lake – Rabbit Lake areas.

Basement rock compositions and structures (Fig. 4B) likely influenced the alteration mineral chemistry of the overlying sandstone. One apparent spatial association is the above-described regional illite (+chlorite+dravite) anomaly that overlies a 5 to 20 km wide aeromagnetic low, where underlying Wollaston Supergroup gneiss includes abundant metaquartzite and metapelite units. The illite anomaly is expressed as local K anomalies in ternary K-U-Th ground spectral gamma-ray surveys (e.g. Shives et al., 2000) but is not evident in regional airborne gamma-ray data (Carson et al., 2002b). Chlorite alteration dominates in the eastern part of this illite region, possibly spatially associated with quartzite ridges, although not all known basement ridges are overlain by this alteration.

A third district-scale alteration effect is subtly evident in airborne radiometric, borehole geophysical, stratigraphic, and mineralogical data. Thorium anomalies correlate directly

with increased detrital grain size in the Manitou Falls Formation, but U and K appear to be systematically depleted with respect to Th and other elements irrespective of grain size (Mwenifumbo et al., 2007). These correlations are subtle in the western part of the Athabasca Basin at Shea Creek (Mwenifumbo et al., 2000) but are strongly developed in conglomerate and pebbly sandstone beds of the Manitou Falls Formation in the eastern part of the Basin. Thorium anomalies in coarse-grained beds are evident in all drill logs in transects across exploration sites (Mwenifumbo and Bernius 2007, Yeo et al., 2007b). Thorium is also anomalous in regional airborne spectral gamma-ray data that map Th anomalies in ternary K-U-Th plots (Carson et al., 2002b), coinciding with the outcrop distribution of the Bird and lower Collins members of Manitou Falls Formation (Campbell et al., 2002, 2007). Detailed correlation of Th anomalies in drill core indicates aluminum phosphate host minerals (Mwenifumbo and Bernius, 2007). These are interpreted as alteration products of heavy minerals, such as monazite, that were preferentially altered in situ to aluminum phosphate minerals and released U (Jefferson et al., 2007; Mwenifumbo et al., 2007).

#### Ages of Known and Prospective Districts

Sedimentation began in the eastern Athabasca Basin at about 1740 to 1730 Ma (Orrell et al., 1999; Rainbird et al., 2007) and slightly earlier in the west while the Wollaston fold and thrust belt was still uplifted (Ramaekers et al., 2007). The Barrenland Group of Thelon Basin also has a minimum age of 1750 to 1720 Ma (Miller et al., 1989; Rainbird et al., 2003a). The uppermost ages of the Athabasca and Barrenland groups are weakly constrained. The Wolverine Point Formation in sequence 3 of the Athabasca Group was deposited at  $1644 \pm 13$  Ma (Rainbird et al., 2007). Carbonaceous marine shale of the Douglas Formation was deposited approximately 100 Ma later at  $1541 \pm 13$  Ma (Creaser and Stasiuk, 2007), but there are no age constraints for the dolomitic Carswell Formation at the top of sequence 4, and no other ages in the Thelon or Hornby Bay basins.

Fluorapatite ages of 1640 to 1620 Ma (Rainbird et al., 2003b) in the Athabasca Basin suggest a regional hydrothermal event at about the same time as localized pre-ore alteration minerals developed (1670-1620 Ma; Alexandre et al., 2003). Athabasca Basin U deposits record two primary hydrothermal ore-related events: 1600 to 1500 Ma (Alexandre et al., 2003) and 1460 to 1350 Ma (McGill et al.,

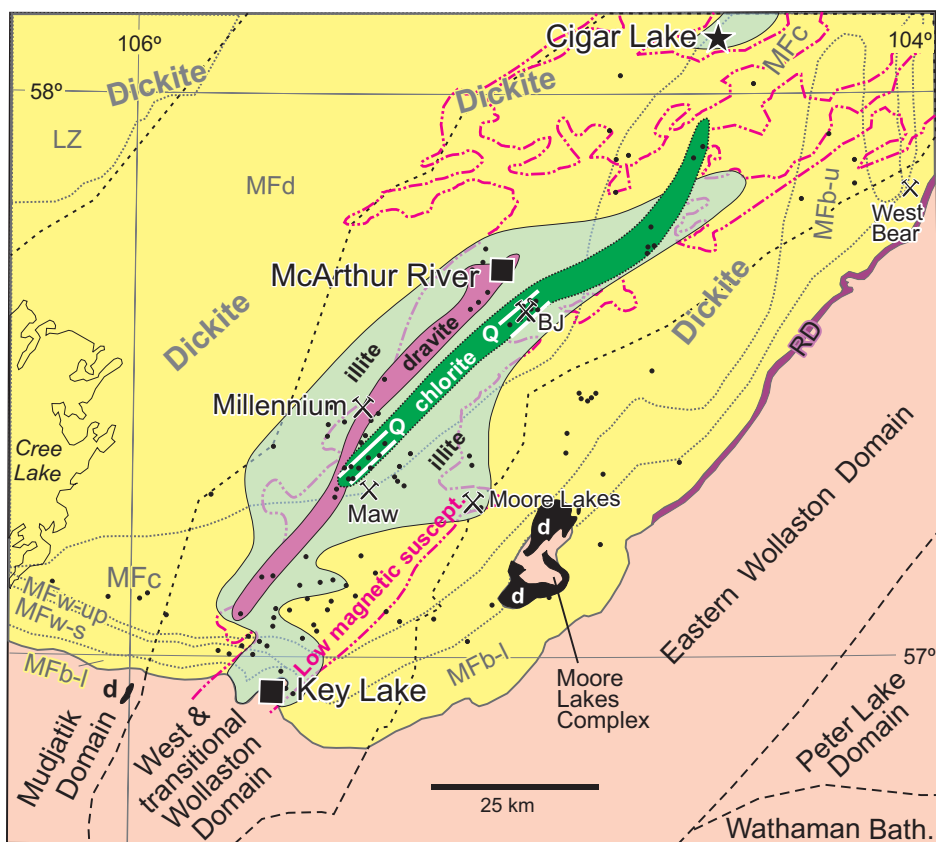


FIGURE 7. Lithogeochemical map of the southeastern part of the Athabasca Basin, showing regional illite, chlorite, and dravite anomalies in the surficial material and outcrops of the Athabasca Group. Also projected to surface are areas of low magnetic susceptibility in basement rocks, and basement quartzite ridges (white bars labelled Q; after Earle and Sopuck (1989)). Basement units (long dash) are projected (short dash) beneath the Athabasca Group (grey) based on aeromagnetic (Fig. 4B) and drill data (dots). RD=Read Formation. Stratigraphic units in the Manitou Falls Formation are MFb-l and MFb-u, lower and upper subunits of Bird Member; MFw-s and MFw-up, sandy and upper pebbly subunits of Warnes Member; MFc, Collins Member; and MFd, Dunlop Member. LZ=Lazenby Lake Formation. d=diabase; square=mine or mill; star=planned mine; crossed hammer=prospect.

1993; Fayek et al., 2002a). These were overprinted by further alteration and U remobilization events at approximately 1176, 900, and 300 Ma (Hoeve and Quirt 1984; Cumming and Krstic 1992; Kyser et al., 2000; Fayek et al., 2002a). Thus, U deposits began to form at the base of the Athabasca Group after the Wolverine Point was deposited 1000 m above, but before the Douglas Formation was deposited an additional 500 m above that. Mineralization therefore took place beneath 1000 to 1500 m of strata, after early diagenesis and during late, high-temperature diagenesis within a time span of at least 100 Ma. Ages of U and associated alteration minerals in northern Australia have similar punctuated histories following primary uraninite deposition at about 1680 Ma (Fig. 6) temporally linked to hydrothermal events recorded in the overlying Kombolgie Subgroup (Polito et al., 2005).

Unconformity-associated U deposits may have formed beneath the Thelon Basin at about the same time as the Athabasca Basin (Kyser et al., 2000), although the oldest date is 1400 Ma (Kiggavik deposit, Fuchs and Hilger, 1989). The Hornby Bay and Elu basins began forming at about the same time as the Thelon and Athabasca basins, also experiencing volcanism at about 1670 Ma (Narakay volcanic complex, Bowring and Ross, 1985), but no unconformity U occurrences have been dated in them.

Pitchblende veins in the Beaverlodge district yield uraninite ages of about 1780 Ma (Koeppel, 1967) near the unconformity between gneiss and the Martin Group. Mafic flows in the upper Martin Group (Gillies Channel Formation) are interpreted to be ca. 1820 Ma based on the age of a diabase dyke with similar geochemistry (Ashton et al., 2004). The Baker Lake Group (1833-1785 Ma, Rainbird et al., 2007) is correlative with the Martin Group (Donaldson, 1968; Fraser et al., 1970; Rainbird et al., 2003a; Ashton et al., 2004) but it differs in having significant intercalated ultrapotassic volcanic rocks (Peterson et al., 2002). Its U prospects appear to be spatially associated with the basal unconformity (Miller, 1979, 1980), although only a few attributes support the unconformity model (Gandhi, 2007).

In the Paleoproterozoic Otish Group of Quebec (Chown and Caty, 1973), unconformity-associated U prospects (Johan et al., 1987) are of unknown age (Ruzicka, 1996a). The Sims Formation, a sequence of conglomerate, arkose, and quartzite of ca 1.8 to 1.45 Ga age located in Labrador (Ware and Hiscott, 1985; Wardle, 2005), unconformably overlies deformed metasedimentary rocks of the Paleoproterozoic Labrador Trough and is being explored for unconformity-associated U (e.g. Consolidated Abaddon Resources, Inc., news release, Feb. 23, 2006). Supracrustal units in the Central Mineral Belt of Labrador may include redbeds and unconformities, however in this region the U deposits with reasonably assured resources, such as Michelin, have been classified as volcanic (Gandhi, 1996).

Older Paleoproterozoic redbed sequences, such as the <1.96 Ga upper Hurwitz Group (Davis et al., 2005), the upper Wollaston Supergroup (Yeo and Delaney, 2007), and upper Huronian Cobalt Group, experienced different atmospheric and tectonic environments. Their dark, locally phosphatic metapelitic units contain regionally elevated U contents; their red quartzite units accumulated in more tectonically active paralic and foreland basin environments and most were stable for only tens of Ma. The Wollaston Supergroup and Hurwitz Group have been variably tectonized and dismembered, forming parts of the basement assemblages beneath the Athabasca and Thelon basins. The Cobalt Group is flat-lying and well enough preserved that unconformity-associated U attributes, such as clay alteration, should be recognizable.

Toward the upper part of the Mesoproterozoic era, the Sibley Group filled the intracontinental Nipigon Basin between 1537  $\pm$  10/-2 Ma, (Davis and Sutcliffe, 1985) and 1339  $\pm$  33 Ma (Franklin, 1978). It is 900 m thick (Hollings et al., 2004), relatively flat lying, fluvial-lacustrine (Cheadle, 1986), and aeolian (Rogala 2003), and is a target of exploration (e.g. Rampart Ventures Ltd., news release, September 30, 2004), despite being arkosic, carbonate-cemented, and lacking alteration in the areas sampled (Hanley et al., 2003). It does record synsedimentary faulting (Craven et al. 2007b), is underlain by regolith, and some of the underlying basement rocks have high background U contents, graphitic metasedimentary units, and local pitchblende veins.

### *Deposit-Scale Geological Attributes*

#### Local Geological Settings

The relationship between reactivated faults and U deposits was illustrated at the deposit scale by Ruhrmann

(1987) for the Gaertner orebody at Key Lake, McGill et al. (1993) for the McArthur River area, Baudemont and Federowich (1996) for the Dominique-Peter deposit, Baudemont and Pacquet (1996) for the McClean Lake area, Rippert et al., (2000) for Shea Creek, Harvey and Bethune (2007) for the Deilmann orebody at Key Lake, and Tourigny et al. (2007) for the Sue C mine.

Drill core of the McClean Lake area preserves evidence of repeated episodic brittle fault reactivation (Baudemont and Paquet, 1996). The geometries of ore zones and reactivation structures in the Sue C Pit at McClean Lake (Tourigny et al. 2007) demonstrate spatial and geometric relationships between transpressional reactivation of ductile Hudsonian basement fault zones and deposition of uraninite in dilatant jogs, consistent with the overall south-plunging deposit geometry.

In the McArthur River area, structural and stratigraphic analysis of drill core (McGill et al. 1993) documented 40 to 80 m of reverse offset along the P2 fault that hosts the ore pods. The ore pods are localized where cross-faults intersect the P2 fault (Györfi et al., 2007). At Key Lake, McClean Lake, and McArthur River faults that were active during sedimentation (Bernier, 2004; Long, 2007, Yeo et al., 2007) appear to have been reactivated after lithification and early silicification, and were conduits for hydrothermal fluid flow in the vicinity of ore deposits (e.g. Hoeve and Quirt, 1984). The spatial linkage between synsedimentary faults and ore-related fault reactivation is not universal, for example at Cigar Lake there is little evidence of offset along graphitic basement units that underlie the ore zone.

Pre- and syn-Athabasca Group paleotopographic features, such as paleovalleys up to a few dekametres deep and minor fault scarps in the order of metres high, have been documented by detailed sedimentological, isopach, and stratigraphic analysis of the Deilmann Pit (Collier and Yeo, 2001; Harvey and Bethune, 2007; Long, 2007) and of the Sue C Pit (Long, 2001, 2007). The concept of paleovalleys spatially associated with Athabasca ore deposits was introduced by Wallis et al. (1985) or possibly earlier, and Macdonald (1980) showed that metapelite and meta-semipelite units were prone to deeper regolith development, especially along graphitic zones. However, recent EXTECH IV work has clarified long-term temporal and spatial relationships between sedimentation, basement paleotopographic lows, and reactivated fault zones (Ramaekers, 2004, Ramaekers et al., 2007). These are related specifically to the faults that host the U deposits and are the focus of alteration halos that can be mapped by various methods, such as mineralogy, seismic reflection, magnetotellurics, and gravity.

Subtle basement uplifts developed before and during sedimentation, and grew to hundreds of metres after sedimentation (Jefferson et al., 2001; Györfi et al., 2002, 2007; Bernier, 2004; Ramaekers et al., 2007; Yeo et al., 2007b) in the McArthur River camp, the BJ zone east of the P2 fault, and in a parallel transect across the Wheeler River area to the south, just east of the Millennium deposit. These commonly termed 'quartzite ridges' (Earle and Sopuck, 1989; Fig. 5 of Marlatt et al., 1992) are mainly compressional pop-up structures bounded by outward-divergent faults (Györfi, 2006; Györfi et al., 2007) such as the hanging wall of the P2 fault that hosts the McArthur River U deposit. Faults exposed

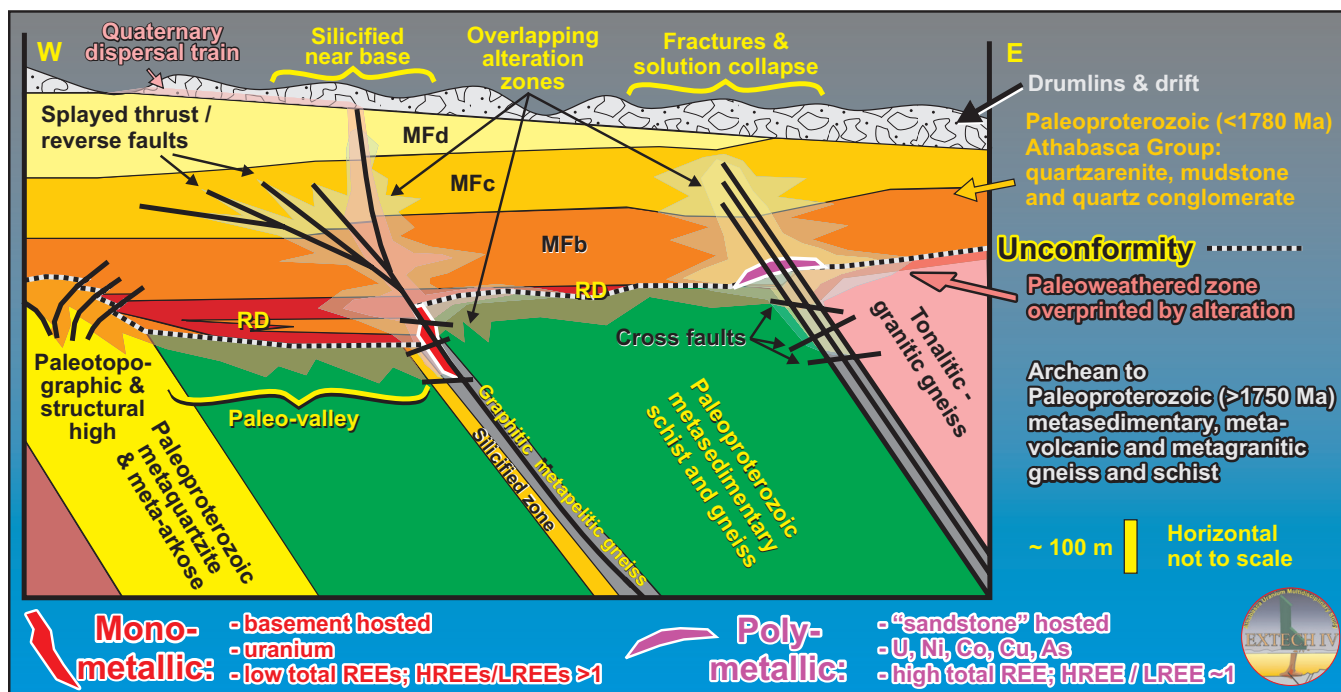


FIGURE 8. Generalized geological elements of mono- and polymetallic unconformity-associated uranium deposits in the eastern part of the Paleoproterozoic Athabasca Basin (after Sibbald et al. (1976), Hoeve and Sibbald (1978), Hoeve and Quirt (1984), McGill et al. (1993), Ruzicka (1996a), Thomas et al. (2000), and Tourigny et al. (2007)). This is an empirical geological framework model illustrating two end-member styles of ore exemplified by the McArthur River (simple) and Cigar Lake (complex) deposits, between which a complete spectrum of styles are known, even within single deposits and deposit groups.

near the unconformity (e.g. Sue C Pit, Tourigny, 2007) splay out into kink folds and bedding-parallel shears in the Manitou Falls Formation. Basal paleotalus deposits, up to 20 m thick, demonstrate that some of the basement uplifts developed before and during deposition, but structural draping of thick tabular units, such as the Read Formation over such basement highs, also records post-depositional uplift.

The graphitic metapelitic gneiss units are also strong conductors, serving as excellent exploration targets for electromagnetic methods in the Athabasca and Thelon basins of Canada and the Kombolgie Basin of Australia. They are also the continuing topic of genetic debate (discussed below under *Focus of Uranium Deposition*).

#### Deposit Size, Morphology, and Architecture

Deposit tonnages and grades summarized in Appendix 1 and Figure 3 are aggregates for ore zones, ranging from a series of lenses to an individual open-pit mine, such as the Sue C, or several large pods, such as the underground McArthur River mine. The details of individual deposit morphology are highly varied, ranging between end-member styles that reflect both stratigraphic and structural control (Hoeve and Quirt 1984; Sibbald 1985; updated in Thomas et al., 2000) (Figs. 8, 9, 10, 11):

1. fracture-controlled and breccia-hosted replacement, dominantly basement-hosted (Fig. 9C; e.g. McArthur River, Rabbit Lake, Eagle Point, McClean-Sue C, Dominique-Peter, Raven and Horseshoe), and
2. clay-bounded, massive ore developed along the unconformity and just above it in the overlying conglomerate and sandstone of the Athabasca Group (Fig. 9A; e.g. Cigar Lake, Key Lake (Deilmann and Gaertner zones),

Collins Bay A, B, and D zones, other McClean deposits, Midwest deposit, and Cluff Lake D zone).

The fracture-controlled basement ore typically occupies steeply to moderately dipping brittle shear, fracture, and breccia zones hundreds of metres in strike length that extend down-dip for tens to 400 m into basement rocks below the unconformity (e.g. Eagle Point, Fig. 9C). Disseminated and massive uraninite/pitchblende occupies fractures and breccia matrix. The high-grade ore lenses are bounded by sheared and brecciated graphitic schist that contains smaller lenses of similar material, forming an envelope of lower grade ore. Typical mining grades for these deposits are on the order of 0.5 to 2% U. Individual lenses of high-grade ore range from 1 to 2 m thick and 3 to 5 m in vertical dimension (e.g. Sue Pit, Tourigny et al., 2002) to massive pods 100 m or more in vertical extent, 90 m in length and 50 m in width, with mining grades in the order of 20 to 25% U at the world-class McArthur River deposit (Jamieson and Spross, 2000).

In contrast, clay-bound ore is developed along the basement-sandstone unconformity and forms flattened elongate pods to flattened linear orebodies typically characterized by a high-grade core (1-15%  $U_3O_8$ ) surrounded by a lower grade halo (<1%  $U_3O_8$ ). The largest of these, Cigar Lake, is about 1.9 km long and 50 to 100 m wide with an upward-convex lens-shaped cross-section up to 20 m in thickness (Andrade, 2002). Most of the clay-bound orebodies have root-like extensions into the basement, resulting in elongated, skewed, T-shaped cross-sections (Fig. 9B). In places, U also extends up into the overlying conglomeratic sandstone, along cataclastic breccia and fracture zones. Isolated, small, 'perched' occurrences of disseminated pitchblende are rarely of ore grade but are good indicators of potential ore at depth. These typically are speculated to represent 'young'

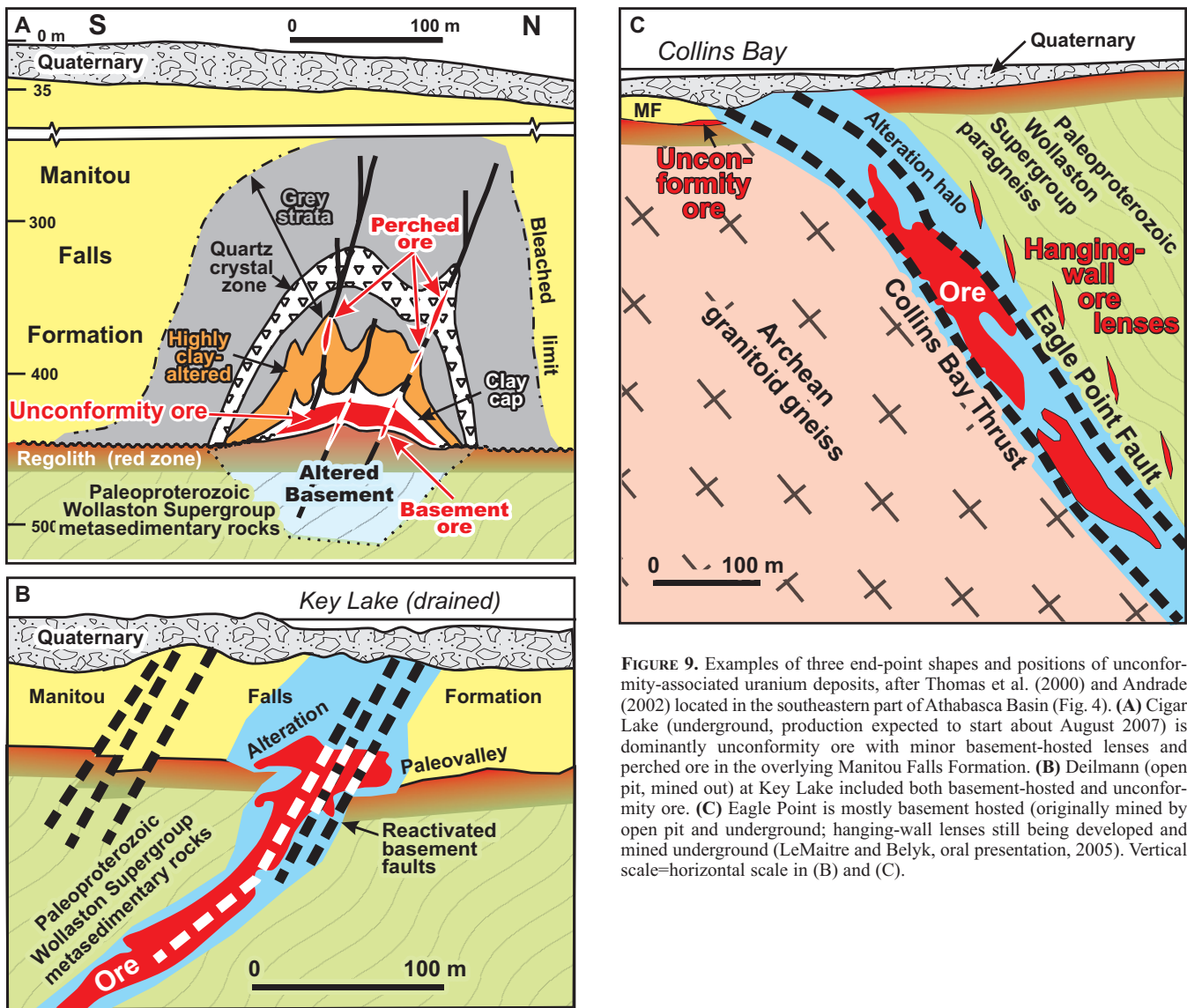


FIGURE 9. Examples of three end-point shapes and positions of unconformity-associated uranium deposits, after Thomas et al. (2000) and Andrade (2002) located in the southeastern part of Athabasca Basin (Fig. 4). (A) Cigar Lake (underground, production expected to start about August 2007) is dominantly unconformity ore with minor basement-hosted lenses and perched ore in the overlying Manitou Falls Formation. (B) Deilmann (open pit, mined out) at Key Lake included both basement-hosted and unconformity ore. (C) Eagle Point is mostly basement hosted (originally mined by open pit and underground; hanging-wall lenses still being developed and mined underground (LeMaitre and Belyk, oral presentation, 2005). Vertical scale=horizontal scale in (B) and (C).

remobilized primary ore. Recent significant intersections in the Shea Creek area (e.g. 27.4%  $U_3O_8$  over 8.8 metres, *Canadian Mining Journal*, July 13, 2005) show that perched ore is a viable exploration target in its own right.

Unconformity-hosted ore deposits show similar ranges in size to fault-hosted ores but in the horizontal dimension. The Cigar Lake deposit overall contains the same order of magnitude of U as McArthur River deposit (Appendix 1), and both deposits comprise major lenses, but those of McArthur River are much more distinct. Only the eastern two lenses at Cigar Lake, with a combined strike length of about 600 m, are scheduled for Phase 1 production, which is estimated to be 496,780 tonnes at an average grade of 20.7%  $U_3O_8$  (Andrade, 2002).

Ore Mineralogy, Chemistry, and Zonation

Unconformity-associated U deposits are dominated by massive to disseminated uraninite (Ruzicka, 1996a). Associated, paragenetically younger, minor coffinite, variable quantities of secondary U minerals, trace to minor sulphide minerals such as galena, pyrite, arsenopyrite, pent-

landite, and chalcopyrite, and native gold characterize the varied metal endowment of these deposits (see below). The field term ‘pitchblende’ is used to refer to the commonly sooty, cryptocrystalline, botryoidal form of uraninite. The sooty appearance of pitchblende is, in part, due to crushing, milling, recrystallization, hydrothermal alteration, and remobilization associated with multistage syn- and post-ore deformation. Much of the ore preserves coarsely crystalline forms of uraninite, and systematic petrographic study of deposits across the Athabasca Basin has revealed paragenetic sequences (e.g. Wilson et al., 2007) that are consistent with classic studies of the Deilmann orebody at Key Lake (Ruhmann, 1987).

The compositional spectrum of unconformity-associated U deposits can be described in terms of monometallic (also known as simple) and polymetallic (complex) end-members on the basis of associated metals (Everhart and Wright, 1953; Beck, 1969; Ruzicka 1989, 1996a; Thomas et al., 2000; Fig. 8). Polymetallic deposits are typically hosted by sandstone and conglomerate, situated within 25 to 50 m of the basement unconformity. At Cigar Lake, nearly all of the

ore is located at the contact with or just above an assemblage consisting of hydrothermally altered paleoregolith and basal sandstone-conglomerate (Andrade, 2000). Polymetallic ores are characterized by anomalous concentrations of sulphide and arsenide minerals containing significant amounts of Ni, Co, Cu, Pb, Zn, and Mo. Some deposits also contain elevated Au, Ag, Se, and platinum-group elements.

Monometallic ores contain just traces of metals other than U and Cu, and are also termed 'simple'. These generally are hosted in basement fractures and faults, typically more than 50 m below the unconformity, with some lenses perched in sandstone well above the unconformity. Eagle Point is an end-member example of a monometallic, entirely basement-hosted U deposit (Thomas et al., 2000). McArthur River exemplifies a super high-grade monometallic deposit that extends from about 20 m above the unconformity (~500 m below surface) to more than 90 metres below the unconformity (McGill et al., 1993; Thomas et al., 2000; Jefferson et al., 2002b). It also contains minor galena, pyrite, chalcopyrite, Ni-Co sulpharsenides, and gold (Gandhi, 2007). A number of deposits have both monometallic and polymetallic components, e.g., Dielmann and Gaertner orebodies at Key Lake (Thomas et al., 2000) that combine to form skewed T-shaped profiles.

#### Alteration Mineralogy and Geochemistry

Alteration mineralogy and geochemistry of unconformity deposits and their host rocks are among the most important exploration criteria in the Athabasca and Thelon basins of Canada and Kombolgie Basin of Australia. The parageneses of these basins (Fig. 6) have been compared by Hoeve and Quirt (1984), Miller and LeCheminant (1985), Kotzer and Kyser (1995), Kyser et al. (2000), and Cuney et al. (2003). Early work on alteration mineralogy in the Athabasca Basin is exemplified by Hoeve et al. (1981a, b), and Hoeve and Quirt (1984). Wasyluk (2002) set the modern template for exploration using clay mineralogy. Intense clay alteration zones surrounding deposits such as Cigar Lake also constitute natural geological barriers to U migration in ground waters (Percival et al., 1993) and are important geotechnical factors in mining and ore processing (Andrade, 2002).

Comparing the mineralogical and fluid paragenesis of the various host basins helps to assess which parameters might be critical for exploration programs. Differences in local to extensive alteration in different basins may suggest different prospectivity; nevertheless, each alteration system must be understood to design the appropriate exploration strategy. District- and corridor-scale high-temperature diagenesis and hydrothermal alteration involving dickite, illite, dravite, and chlorite are described above for the Athabasca Basin. Each alteration product has detailed, deposit-specific distributions that need to be mapped in three dimensions for each exploration target. The original kaolinitic (detrital) clay is locally preserved in early pre-ore (Q1) silicification, such as at McArthur River, along with dickite, its alteration product (Wasyluk, 2002; Mwenifumbo et al., 2007). Such relict minerals need to be recognized in order to determine what is anomalous.

Phosphate minerals preserve a rich record of regional to local, low- to high-temperature saline fluid diagenesis

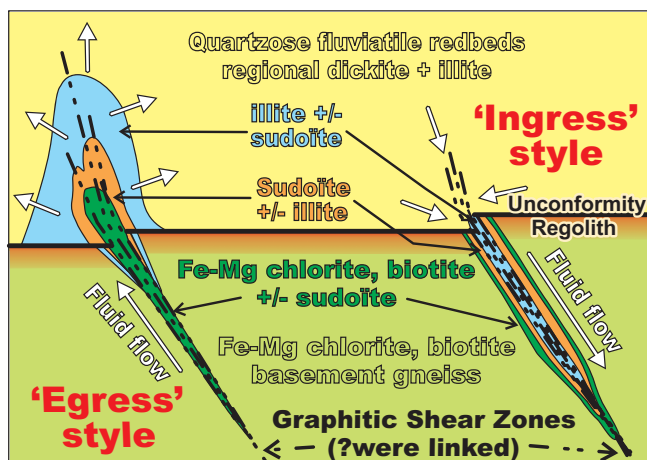


FIGURE 10. Diagrammatic explanation of 'egress'- versus 'ingress'-style alteration zones for unconformity-associated uranium deposits, after Hoeve and Quirt (1984), Sibbald (1985), Fayek and Kyser (1997), and Quirt (2003). Fluids flowing from the basement were reducing; fluids from the redbeds were oxidizing and contained uranium. Models invoke mixing of such fluids at the unconformity to precipitate uranium. The shear zones and fluid flow along them may have been linked.

involving precipitation, destruction, and re-precipitation of phosphate as xenotime, apatite, and Ca-Sr-LREE-Al-phosphate minerals (AP). Xenotime in the Athabasca Basin typically forms 1 to 10 micron euhedral overgrowths on detrital zircon, lacks U, and is overgrown by quartz and fluorapatite (Rainbird et al., 2003b). This xenotime is significantly post-dated by hydrothermal uraniferous xenotime described by Quirt et al. (1991) for the Maw Zone (MacDougall, 1990). Diagenetic, variably uraniferous, patchy to stratabound fluorapatite cement was dated at ca. 1640 to 1620 Ma by U-Pb SHRIMP geochronology from the Wolverine Point to basal Fair Point formations (Rainbird et al., 2003b). This age is within error of the depositional age of the Wolverine Point Formation ( $1644 \pm 13$  Ma, Rainbird et al., 2007). Fluorapatite in the Thelon Basin has a similar paragenesis, locally forming patchy pinkish red cement with approximately 20 to 500 ppm U (Miller, 1983), dated as 1750 to 1720 Ma (Miller et al., 1989), and spatially associated with fracture and breccia zones from the bottom to top of the Thelon Formation.

The pseudocubic AP minerals recognized in trace to minor amounts throughout the Athabasca Group are interpreted as mid- to late-diagenetic to early hydrothermal. Goyazite, intergrown with illite and dravite (Hoeve and Quirt, 1984) was estimated at 1500 to 1250 Ma in age by Kotzer and Kyser (1995). Crandallite, in the same family as goyazite, has been interpreted as the youngest phosphate generation, intermixed with late-formed kaolinite (Hoeve and Quirt, 1984, Wilson, 1985). The AP minerals referred to as 'crandallite group' by Mwenifumbo and Bernius (2007) were specifically identified as florencite, form abundant 3 to 6 micron grains intimately intergrown with illite, dickite, anatase, and hematite, and are most abundant in the lower Manitou Falls Formation around the eastern Athabasca Basin. They contain elevated Th, form clusters resembling detrital grains, and are the matrix to hematite-rich pebbly laminae with relatively abundant detrital zircon and rare xenotime. This assemblage is interpreted as altered from



detrital heavy minerals during moderately high-temperature (150–170°C) diagenesis that post-dated Q1 quartz overgrowths. In the Thelon, Hornby Bay, and Elu basins, aluminum phosphate sulphate (APS) minerals are described as concentrated at the base and in the regolith (Gall and Donaldson, 2006). These APS are also pseudocubic but slightly larger than the Athabasca AP minerals, and here too appear to post-date uraniumiferous fluorapatite.

Other alteration is spatially associated with ore deposits located inside and outside the above-described regional zones. Anomalously high proportions of illite are observed in the 1 to 3% clay matrix of the Athabasca Group strata and in altered basement rock in the vicinity of U deposits. This results in anomalous  $K_2O/Al_2O_3$  ratios in the sandstone (Earle and Sopuck, 1989) that are locally recorded by ground and airborne spectral K-Th-U gamma-ray and till geochemical data (Shives et al., 2000; Campbell et al., 2007). Sudoite, an Al-Mg-rich ditrioctahedral chlorite, is present in the alteration of both sandstone and basement at Cigar Lake (Percival and Kodama, 1989) and McArthur River (Wasyluk, 2002). Up to five types of chlorite have been documented in the basement (Quirt and Wasyluk, 1997) – the challenge for chlorite is to distinguish ore-related alteration from that associated with retrograde metamorphism and paleoweathering.

Quartz cement predating ore formation (Q1) is informally termed ‘tombstone’ where drill core has a polished appearance due to a high density of cement in Athabasca Group strata above or proximal to basement quartzite highs above and to the west of the McArthur River area and in the Millennium area (Yeo et al., 2001b; Mwenifumbo et al., 2004, 2007). The tombstone silicification of the Athabasca Group preserves some of the early detrital kaolinitic clay mineralogy, detrital to early diagenetic hematite and the early regional diagenetic dickite (Wasyluk, 2002). Even some microbial laminae defined by very finely crystalline hematite are preserved by silicification (Yeo et al., 2007a) very close to the McArthur River deposit.

Quartz dissolution is a major alteration effect above the Cigar Lake deposit, resulting in significant volume reduction and collapse of the Athabasca Group strata. It is superimposed in places on Q1 at McArthur River. Later silicification fronts comprising drusy disseminated to fracture-filling quartz (Q2) are also present in both larger quartz-dissolution alteration systems, for example at Cigar Lake (Andrade 2002) and in the McArthur River area (McGill et al. 1993). Drusy quartz (quartz crystals filling void space) is mostly developed at the periphery of the ore deposits, is related to quartz dissolution in the deposit area by mass balance analysis (Percival, 1989), and was probably synchronous with quartz dissolution during deposit formation (Hoeve and

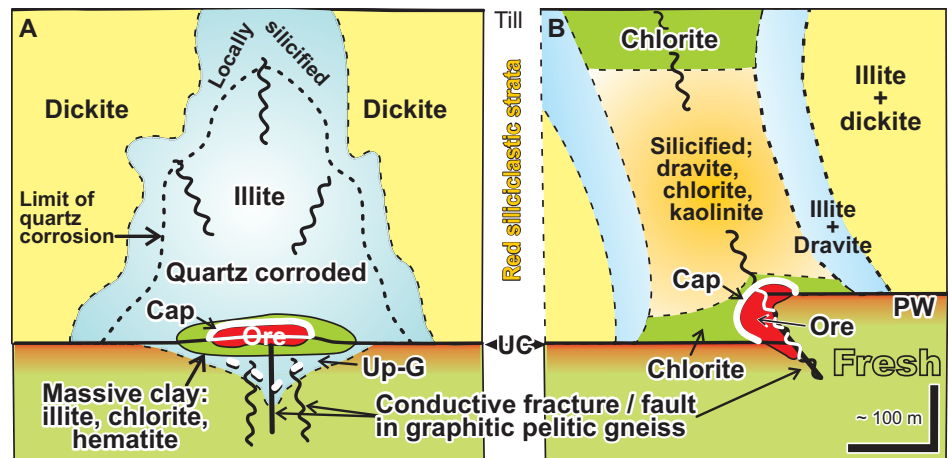


FIGURE 11. Two end-member ‘egress-type’ sandstone alteration models for uranium deposits in the eastern Athabasca Basin (modified from Matthews et al. (1997), Wasyluk (2002), and Quirt (2003)). (A) Quartz dissolution egress style (e.g. Cigar Lake, Midwest). (B) Silicification egress style (e.g. McArthur River, Key Lake). Cap=secondary black-red earthy hematite capping ore bodies, Reg=regolith profile grading from red hematitic saprolite down through green chloritic alteration to fresh basement gneiss, UC=unconformity, Up-G=upper limit of graphite preserved in footwall alteration zone. Fresh=unaltered metamorphic basement: Fe-Mg-chlorite biotite paragneiss. Note that the silicification is a very early component of the right-hand model, and that quartz dissolution is also present there, albeit a minor localized feature.

Quirt 1984). Later drusy quartz (Q3) was also described locally within the previous quartz-dissolution zones (Thomas et al., 2000; Andrade, 2002).

The alteration types described above are organized into two broad geometric shapes interpreted as ‘egress’ and ‘ingress’ (Fig. 10). The egress and ingress alteration zones are spatially but not necessarily temporally related to ore deposits. Egress-type alteration halos are developed mainly in the conglomeratic sandstone overlying unconformity-associated U deposits (Hoeve and Quirt, 1984). Deposits with egress halos include both basement-hosted and sandstone-hosted types, and the alteration ranges between two distinctive end-member types as illustrated in Figure 11: i) quartz dissolution + illite, and ii) silicified (Q1 +Q2) + later illite-kaolinite-chlorite + dravite. Strata overlying deposits in the northern part of the eastern Athabasca Basin characteristically underwent quartz corrosion with volume losses locally exceeding 90% (Percival, 1989). In contrast, alteration in the McArthur River to Millennium corridor is dominantly represented by the early silicification end-member with local, intense, pre-ore quartz corrosion and little apparent volume loss (Matthews et al., 1997). Around the Deilmann orebody at Key Lake, silicification is minor but late kaolinite and dravite (tourmaline) are superimposed on earlier dickite alteration.

Illite-kaolinite-chlorite alteration halos are up to 400 m wide at the base of the Athabasca Group (Figs. 10, 11), thousands of metres in strike length, and extend several hundred metres above major deposits (e.g. Cigar Lake, Bruneton, 1993; McArthur River, Thomas et al., 2000; Shea Creek, Kister et al., 2003). This alteration typically envelops the main ore-controlling structures, forming plume-shaped or flattened elongate bell-shaped halos that taper gradually upward from the base of the sandstone and narrow sharply downward into the basement. Illite-dominated halos have  $K_2O/Al_2O_3$  ratios  $>0.18$  and  $MgO/Al_2O_3$  ratios  $<0.15$ ; kaolinite-dominated halos have  $K_2O/Al_2O_3$  and  $MgO/Al_2O_3$  ratios  $<0.04$ ; and chlorite-rich haloes have  $MgO/Al_2O_3$

ratios  $>0.125$  and  $K_2O/Al_2O_3$  ratios  $<\sim 0.04$  (Sopuck et al., 1983). Compared to background ratios of 0.1 to 0.16  $K_2O/Al_2O_3$  in the Athabasca Group (Ibrahim and Wu, 1985), Percival (1989) measured  $K_2O/Al_2O_3$  ratios  $>0.27\%$  for most of the alteration zone at Cigar Lake ( $n=150$ ).

A bleached and white clay (illite) replacement zone at the top of the unconformity is interpreted as ore-related hydrothermal alteration superimposed on the lateritic-weathered, red-green profile developed in basement rocks (Macdonald, 1980). The bleached white to pale green colour also locally overprints very fine-grained early diagenetic hematite in red to pink mudstone beds and intraclasts of the Read Formation. Intervening sandstone and conglomerate beds in the alteration zones are bleached to very pale green or white. Early diagenetic hematite is also intimately inter-layered with clay minerals to form micro-laminae in oncoidal beds that mark the base of the Smart Formation (Yeo et al., 2007a). The matrix of these oncoidal beds is also intensely hematitic, deep maroon in colour, and very fine grained. These oncoidal beds are suspected to have resulted from pedogenic and/or shallow-water processes in ephemeral ponds. Early microbial laminae outlined by microcrystalline hematite are preserved by Q1 silicification within a bleached zone near McArthur River.

Iron pigmentation takes a number of additional forms in the Athabasca Basin. Away from ore zones, late diagenetic, broadly developed hematite transects sandstone and conglomerate beds, is commonly purple to maroon in colour, and forms spots or liesegang bands with roll shapes that tend to follow bedding planes. Bleached zones transecting this late diagenetic hematite are purplish to reddish brown, through very pale purple to nearly white in colour. Bright to very dark, 'brick red', coarse-grained hematite forms caps over ore deposits (Fig. 11). Brick red through deep reddish brown to nearly black hematite also forms dense cement within parts of the Read Formation and Bird Member of Manitou Falls Formation (Table 3), particularly in the lower conglomeratic subunits that overlie the basal unconformity and are near U deposits. This very intense, crystalline hematite alteration is interpreted as hydrothermal in origin. Recent oxidation processes are documented by limonitic alteration along fault zones, and by local limonitic liesegang banding in outcrop and drill core.

Less alteration is evident above basement-hosted deposits with ingress-type alteration zoning (Hoeve and Quirt, 1984). Such deposits are essentially 'blind' exploration targets, except for geophysical methods, although broader geochemical and mineralogical halos above them may provide clues to their existence. They are monomineralic and have very narrow, inverted alteration halos along the sides of the basement structure, grading from illite±suoite on the inside, through suoite±illite, to Fe-Mg chlorite±suoite on the outside against fresh basement rock (Fig. 10; Quirt 2003). Targeting these metallurgically attractive deposits requires exploration geologists to understand the geometry of fault systems. Some alteration zones have both ingress and egress characteristics (e.g. McArthur River), suggesting complex hydrothermal systems involving both processes very close to one another in both time and space. Such fluid-flow complexities have previously been interpreted for Canadian and Australian deposits (e.g. Hoeve and Quirt, 1984; Wilde and

Wall, 1987). A good understanding of the basement geology and structural features is required for basement-hosted deposit exploration.

## Key Exploration Criteria

### *Geological Exploration Criteria*

The main first-order exploration criterion is Paleo- to Mesoproterozoic redbed basins as described above under continental- and district-scale geological attributes. The sedimentary sequences in Canadian basins with known U resources are depleted in U. On the other hand, a number of such basins in Canada and around the world have yet to reveal such deposits, although prospects have been discovered. Another first-order criterion is basement complexes characterized by relatively high U, well above the Clarke value of about 5 ppm (Thomas, 1983; Annesley et al., 2005). These are deformed and metamorphosed, tectonically interleaved Archean and Paleoproterozoic orthogneiss and paragneiss, intruded by granitoid plutons and pegmatite bodies.

Second-order empirical parameters associated with unconformity mineralization include graphitic metapelite, ductile faults, and other pre-existing complexities (e.g. extensional or compressional flexures, bifurcations, splays, duplex structures, and cross structures) within the basement complex. Repeated brittle reactivations of the ductile structures offset the basal unconformity, and were foci for fluid flow and ore deposition. Reactivation structures in sandstone can be traced into the primary basement fault zone and provide the local structural framework of a prospect. Reactivated fault zones may be localized at hinge lines that separate different depositional subbasins associated with different sequences. Those flexures that developed before and during sedimentation would have provided the most intensive 'ground preparation' for mineralization. In the case of the Athabasca Basin, the first U ore was emplaced during and after deposition of upper sequence 3 (Fig. 7; see *Ages of Known and Prospective Districts* above).

Paleovalleys and post-depositional offsets of the basal unconformity are manifestations of the heterogeneous nature of the basement rocks described above. Paleovalleys are not a pre-requisite for world-class orebodies; for example, the Cigar Lake deposit rests on a small basement high (Andrade, 2002). Intersections of different arrays of steeply dipping faults are especially significant, such as between the P2 fault and cross faults at McArthur River (e.g. Fig. 4 in McGill et al., 1993; Györfi, 2006). A variety of structural sites favourable for U deposition have been documented at intersections between the Rabbit Lake and a number of other fault trends in the Rabbit Lake – Eagle Point area (Rhys, 2002; LeMaitre and Belyk, oral presentation, Targeted Geoscience Initiative Saskatchewan Open House, 2004; Thomas et al., oral presentation, Targeted Geoscience Initiative Saskatchewan Open House, 2005).

### *Geochemical Exploration Criteria*

Geochemistry in various media reflects the pervasive and local mineralogical alteration. The regional background of U is 1 to 2 ppm in lake sediments (mainly glacial till reflecting local bedrock sources; Maurice et al. 1985) and approximately 1 ppm in the Athabasca Group (Quirt, 1985; Wallis et

al., 1985; Percival, 1989; Andrade 2002, Table 4.1). Uranium anomalies in lake sediments reach values of 1500 ppm in the Key Lake area (Maurice et al., 1985). Anomalous U (>2.5 ppm) in the Athabasca Group was discovered in the above described clay alteration halos, extending in places to the top of the sandstone, even in sections more than 500 m thick (e.g. Clark, 1987; Thomas et al., 2000). Percival (1989) measured common values of 13 ppm in 'unaltered' sandstone above the clay alteration halo at Cigar Lake, with highly altered sandstone in the clay zone yielding up to 235 ppm U and altered basement giving approximately 95 ppm. The trace elements U, Ni, As, and Co have greater than background concentrations in halos above some deposits and prospects. The dispersion of Ni, As, and Co in such geochemical anomalies is restricted in some cases to tens of metres (Sopuck et al., 1983), thus limiting their usefulness as pathfinder elements.

Lake water and sediment geochemistry (e.g. Coker and Dunn, 1983; Maurice et al., 1985) and radiometric prospecting were significant tools in early regional exploration. Measuring and contouring radon gas emission as an expression of radioactive decay related to underlying U ore deposits has been used with mixed results on reconnaissance to detailed scales (e.g. Dyck, 1969; Scott, 1983), but is still employed today. Analysis of spruce twigs showed that the McClean Lake – Rabbit Lake area is situated in the middle of an immense biogeochemical anomaly that was interpreted as a result of tree roots extracting anomalous U from ground water (Dunn, 1983). Groundwater geochemistry has been employed successfully in the past (e.g. Toulhoat and Beaucaire, 1993), however, given the long history of fluid flow and the still-active but variably constrained groundwater systems in the broadly permeable Athabasca Group, this technique should be re-evaluated.

As exploration advanced to deeper targets, focus shifted to alteration mineralogy reflected by surficial geochemistry. Regional-scale alteration halos of potassic clay minerals (e.g. illite), boron minerals (e.g. dravite), quartz cement, and dissolution are intersected in various places at the present surface where they are incorporated into Quaternary till. These in situ to slightly transported anomalies can be measured in till and rock samples (Earle and Sopuck, 1989; Campbell et al., 2007) and by gamma-ray spectrometry as outlined below.

Favourable basins show geochemical evidence of regional to focused fluid flow resulting in mineralogical expressions such as clay alteration and redox boundaries. Illite, chlorite, dravite, quartz cement and dissolution are the main local vectors in ingress-type or expanded egress-type zonation. This mineralogy can be analysed in the field by portable short-wave infrared (SWIR) spectrometers such as PIMA II<sup>©</sup> (Integrated Spectronics Ltd.) and FieldSpec Pro. Calibrated software algorithms for semiquantitative analyses (Earle et al., 1999) enhance the usefulness of these spectrometers. Spectrometric methods have the potential to be fully quantitative, given calibration of peak resolution with appropriate mineral standards, and the use of artificial mixtures to develop best-fit algorithms (Zhang et al., 2001; Percival et al., 2002). Infrared spectrometry is particularly useful in distinguishing between the kaolinite-group polytypes of kaolinite and dickite (Wasyliuk, 2002). Normative calculations

based on lithochemical data further refine the mineralogical identifications.

Airborne gamma-ray spectrometry is here treated as a geochemical tool, because it directly measures U, K, and Th in surficial material. Interpretation of results from such surveys requires knowledge of paleo-ice-flow directions and till stratigraphy. Campbell et al. (2007) have provided calibration data that document relationships between gamma-ray and surficial geochemical data. This provides a quantitative basis for the use of ground (Shives et al., 2000) and airborne gamma-ray multiparameter geophysical surveys (e.g. Richardson, 1983; Campbell et al., 2002) as geochemical prospecting and lithologic mapping tools. The extensive illite alteration corridor between McArthur River and Key Lake (Fig. 7) does not correlate with K in published reconnaissance gamma-ray data (Carson et al., 2002a, b), although detailed ground gamma-ray spectrometry by Shives et al. (2000) suggests that K does correlate with illite alteration in the McArthur River area.

The mineralogy and chemical composition of Quaternary deposits are strongly related to local bedrock. A variety of ice-flow directions must be considered in tracing surficial materials back to their sources. Campbell (2007) provided an overview of Quaternary geology east of the Snowbird Tectonic Zone. Fenton and Pawlowicz (in press - a, b) reviewed surficial geology in map areas NTS 74 E and 74 L that cover most of the Alberta portion of the Athabasca Basin, compiled regional drift thickness and draped the Quaternary geology on principal 3-component imagery of the RADARSAT-1 data. Campbell et al. (2007) mapped detailed relationships between till composition and airborne gamma-ray data that depend on till stratigraphy and the nature of local bedrock. These reviews highlight morphological features such as sand dunes, drumlins, and eskers, and other indicators of the prevailing southwestward regional ice-flow and complex local ice-flow histories.

In the Athabasca Basin region, the bedrock is broadly the basement gneiss or the Athabasca Group conglomeratic sandstone with their varying degrees of alteration. Transport of gneissic material onto the edges of the basin from the northeast (prevailing ice flow) may be the cause of some anomalous linear features (Campbell et al., 2007). Also, the Athabasca Group material has been transported onto the gneissic basement and Paleozoic strata to the southwest, hence anomalies found there could tend to represent a source somewhere up-ice within the Athabasca Basin. This does not rule out the possibility of anomalies derived from outlying basement-hosted U deposits, above which Athabasca Group cover has been totally eroded.

#### *Geophysical Exploration Criteria*

Initial exploration in the Athabasca and similar basins focused on surface expressions of radioactivity associated with near-surface deposits located around the margins of the unconformities. In the Athabasca Basin this included the rim and the uplifted basement pillar of the Carswell meteorite impact structure. Thus initial discoveries included the Rabbit Lake, Key Lake, and Cluff Lake camps (Appendix 1). Detailed follow-up exploration traditionally focused on airborne and ground electromagnetic methods based on recog-

dition of an association between graphitic faults and U at Key Lake (e.g. Matthews et al., 1997). These methods have been and remain the most effective tool to identify the precise location, depth, and characteristics of basement conductors.

Electromagnetic methods also detect ore-related alteration features. High-resolution airborne electromagnetic surveys in Australia have detected shallow but hidden low-resistivity alteration zones and crudely mapped fault offsets of the unconformity (Bisset, 2003). Improved audiomagnetotelluric methods in the McArthur River area of Athabasca Basin have detected deep conductors and shallow alteration zones (Craven et al., 2007; Tuncer et al., in press). Highly altered, clay-rich, quartz-corroded quartzarenite has relatively low resistivity, whereas quartz-rich silicified zones are characterized by high resistivity. Powell et al. (2005) have shown that high resistivity also maps zones of high porosity related to quartz dissolution and fracturing in the Virgin River exploration area (Centennial prospect, Fig. 4A). Detailed multiparameter borehole geophysics has been used to calibrate audiomagnetotelluric data and link them to detailed lithostratigraphic and mineralogical data, especially the resistivity contrasts (Mwenifumbo et al., 2004, 2007).

Airborne magnetic surveys provide the means to extrapolate maps of basement geology from the margins of these Proterozoic basins to their centres (e.g. Pilkington, 1989) with the aid of magnetic susceptibility and related data from outcrop and drill cores that intersect the basement. Card (2006) and Thomas and McHardy (2007) provide modern reviews of this technology and demonstrate its application to the central and eastern Athabasca Basin, respectively. They point out first-order exploration targets, such as faults and favourable basement lithologic units, as mapped by magnetic gradients between Archean gneiss domes and the Wollaston Supergroup (Fig. 4A).

Seismic reflection is a relatively new exploration tool from the mineral industry perspective, although much of our knowledge about the overall depth and shape of the Athabasca and Thelon basins has come from early seismic studies (e.g. Hobson and MacAulay, 1969; Overton, 1977; Suryam, 1981, 1984). Modern seismic reflection provides a continuous structural framework in 2-D and 3-D (White et al., 2007), from near surface to a few kilometres below the unconformity (Györfi et al., 2007) or deep in the crust to Moho (Hajnal et al., 2007) by varying source frequency, shot and geophone spacing and data processing, calibrated with the aid of borehole geophysics (Mwenifumbo et al., 2004). Complete structural sections can be interpreted using local and generic structural analogues (Fig. 5B) to determine fundamental exploration parameters such as the position of and irregularities in the unconformity, and shallow to deep faults.

Ground and airborne gravity can detect alteration zones as negative gravity anomalies (dissolution zones) or positive anomalies (silicified zones), but direct detection of ore deposits is a challenge due to their small dimensions that limit the magnitude of gravity anomalies (Thomas and Wood, 2007). Gravity also provides insights into the geological framework for exploration on both regional and district scales. It is best used in conjunction with multiple other data sets that can help to resolve ambiguities related to factors such as overburden thickness, and bulk densities and dips of deep basement units.

## Genetic/Exploration Models

### *Conventional Models*

The first refereed publication of a geological model for a new class of U deposit called 'unconformity type' was by Hoeve and Sibbald (1978), which built on the work and ideas of many geologists following the initial discoveries at Rabbit Lake in 1968, Cluff Lake in 1969, and Key Lake in 1975, by which time the importance of the unconformity had become evident. Most conventional models employed today in the Athabasca Basin are a combination of empirical, spatially associated attributes that invoke late diagenetic to hydrothermal processes with ore formation being spatially and temporally focused by the reactivation of pre-Athabasca Group structures (e.g. Hoeve et al. 1980; Kotzer and Kyser 1995; Baudemont and Paquet, 1996; Fayek and Kyser, 1997; Thomas et al., 2000). These models suggest that oxidizing, U-bearing, basin fluids heated by geothermal gradient eventually attained 200°C (burial depths of ~5-6 km) at the unconformity and reacted with reducing fluids coming out of reactivated basement shear zones. Uranium precipitated as uraninite in fault zones where reduced and oxidized fluids were mixed. Uraninite filled tension gashes and other structural traps during active faulting, and was repeatedly brecciated while new uraninite precipitated. Ore deposits accumulated where these conditions were focused for very long periods of time (Hoeve and Quirt, 1987), perhaps hundreds of millions of years (Kyser et al., 2000). Zones of inferred fluid mixing are characterized by alteration halos that contain illite, kaolinite, dravite, chlorite, euhedral quartz, and locally, Ni-Co-As-Cu sulphide minerals (Hoeve and Quirt, 1984; Wallis et al., 1985; Kotzer and Kyser, 1995). The latter described the chlorite as Mg-chlorite (=clinocllore). At Cigar Lake, most of the chlorite is a less common Al-Mg variety termed sudoite (Hoeve and Quirt, 1984; Percival and Kodama, 1989). More than one variety of chlorite likely is present in this deposit type (Quirt, 1989, 2003).

Pre-ore to post-ore alteration halos developed around sites of ore deposition where reduced basement fluids circulated upward into the overlying oxidized basin-fluid environment ('egress type' of Fayek and Kyser, 1997). Ingress of basinal fluids downward into the basement developed inverted and condensed alteration zones, mainly in host basement rocks, with a more subtle and/or complex expression in overlying conglomeratic sandstone (Fayek and Kyser, 1997). There are many variations on the ingress and egress alteration themes in the unconformity-associated U deposit model (Quirt and Ramaekers, 2002; Quirt, 2003). Both flow paths may have developed nearby, along the same fault zone or on intersecting faults, especially in structurally complex areas. Hoeve and Quirt (1984, p. 110-114), Wilde and Wall (1987, p. 1167), and Wilde et al. (1989) discussed similar fluid flow concepts for unconformity-associated U deposits in Australia.

Early models by Knipping (1974), revisited by Dahlkamp (1978) and Langford (1978), introduced the role of supergene processes related to pre-Athabasca Group weathering of basement rocks, transport by surface- and groundwater, and deposition within basement rocks under reducing conditions. In the 1970s, deeply buried deposits (i.e. beneath hundreds of metres of the Athabasca Group) were not known,

but it is now clear that even the deposits close to surface near the edge of the Athabasca Basin were formed after at least the lower Athabasca Group was deposited because of geochronology of the U deposits and the Athabasca Group. Even for deposits more than 500 m down, alteration effects reach the surface (e.g. Fig. 8).

A magmatic hydrothermal origin was briefly considered (summarized in Hoeve et al., 1980), but there is no local evidence of magmatism that is coeval with U deposition. While work on the unconformity model progressed, discussion continued as to the source of U being directly from basement rock (e.g. Tremblay, 1982) or from secondary sources including the Athabasca Group (Ruzicka, 1996a).

### *Genetic Models: Advances of the Last Decade*

Significant advances have been made since the discovery of Rabbit Lake more than 35 years ago, but many new questions have arisen and some of the fundamental enigmas of Hoeve and Sibbald (1978) remain. A wide variety of U deposit models was developed more than a decade ago (Dahlkamp, 1993) and these are still in use today (Organization for Economic Co-operation and Development, Nuclear Energy Agency, and the International Atomic Energy, 2004). A combined efficiency of source, transport, and deposition of U are required to form a world-class deposit, and this combination still needs to be better understood (Cuney et al., 2003). Uranium deposits may appear at each step of the geological cycle, from magmatic and fluid fractionation in the deep continental crust (e.g. the Tranomaro pyroxenite, Madagascar and Rössing alaskite, Namibia) to evapotranspiration at the surface (e.g. the Yeleerie calcrete, Australia). However, very high-grade, large-tonnage U deposits have only been discovered in the vicinity of unconformities of Mesoproterozoic age. In the following reviews of source, transport, and deposition of U, there is sufficient diversity in unconformity deposits to also require multiple variants on the main model.

### Uranium Sources

Suggested primary sources of U for the Athabasca and Thelon basins include radiogenic S-type granites and pegmatites (e.g. Thomas, 1983; Madore et al., 2000), metasedimentary terrains with abundant pelite whose U endowment is well above the 5 ppm Clarke value (Thomas, 1983; Miller and LeCheminant, 1985), and pre-existing U concentrations such as those in the Wollaston Supergroup (Delaney, 1993; Yeo and Delaney, 2007), pegmatites (Thomas, 1983) that intrude the Hearne Province (formerly known as the Cree Lake Zone; Lewry and Sibbald, 1979) and the many deposits in the Beaverlodge (Koeppel, 1967; Tremblay, 1968; Ruzicka, 1996b). As in copper provinces of the world, regions relatively well endowed with U due to a high proportion of radiogenic granitoid intrusions (such as in the Wollaston, Mudjatik, and Wathaman domains of the Trans Hudson Orogen and in the Taltson magmatic zone), have a much better chance of generating world-class deposits given favourable subsequent conditions. Thus a particular set of tectonic conditions was responsible for creating the U-rich western Churchill structural province. The McArthur Basin of Australia also overlies U-rich basement terrains and has a similar metallogenic history (Kyser et al., 2000). The U may

have been removed from these primary sources either directly, from the Athabasca Group via detrital heavy minerals, or from the Athabasca Group via detrital clay and hydroxide minerals. Mass balance calculations illustrating the viability of the first two hypotheses are summarized by Jefferson et al. (2007).

How could U have been derived directly from underlying U-rich basement rock (Annesley et al. 1997; Hecht and Cuney, 2000; Madore et al., 2000)? Low-U AP products of altered uraniumiferous monazite have been documented in basement rocks underlying the Athabasca Group (Hecht and Cuney, 2000; Madore et al., 2000; Cuney et al., 2003), in basement alteration zones proximal to U deposits beneath the Kombolgie Basin of Australia (Gaboreau et al., 2003). However, drill core does not indicate sufficiently large volume and permeability of altered basement immediately beneath the Athabasca Basin to generate the amount of U needed to form major deposits (Hoeve and Quirt, 1984). Transitions to unaltered, very tight basement rocks are sharp outside of shear zones (Fig. 10). Deeper alteration along some fault zones may be partly attributed to paleoweathering. On the other hand, deep seismic profiles (Hajnal et al., 2007) suggest that a large volume of basement rock has been disrupted along the P2 fault to considerable depth.

How could U have been derived from primary detrital minerals in the Athabasca Group? Such minerals are essentially absent except for zircon and rare tourmaline; the other heavy minerals could have been incorporated in the original sediment but destroyed by diagenetic to hydrothermal alteration. Regionally, the group contains  $\leq 1$  ppm U despite its proximal detrital source terrains containing 5 to 20 ppm U. Possible original detrital carriers of U include rock fragments and heavy minerals, such as zircon, monazite, and uraninite, that should have been eroded from source terranes but are now absent, except zircon. Detrital U-oxide must have been rare because of the highly oxidizing conditions and lack of organic matter. Detrital ilmenite and magnetite are consistent with the preservation of iron-titanium oxide minerals in the Manitou Falls Formation (Mwenifumbo and Bernius, 2007), however, ilmenite and magnetite do not typically carry significant U. Feldspar is ruled out as a source because it was such a minor component of the Athabasca Group (see *Quartz-Dominated, Uranium-Depleted Strata* above). Zircon (suggested by Kyser et al., 2000) is ruled out because regionally it is fresh with normal U contents (Rayner et al., 2003; Rainbird et al., 2007), and any altered zircon across the Athabasca Basin shows evidence of U uptake, not leaching (Cuney et al., 2003), a characteristic of altered zircon in general (e.g. Rayner et al., 2005).

If there was a detrital mineral to yield U, monazite is favoured by the presence of Th- and REE- rich AP minerals in coarse-grained beds with black bands outlining cross beds and laminae draped over pebbles (e.g. Fig. 5 of Yeo et al., 2000). In situ alteration of monazite has been shown to produce U-poor but Th- and REE-rich AP minerals in the basement (above), in sandstone of the Franceville Basin, and around the Oklo deposits in Gabon (Mathieu et al., 2000; Cuney and Mathieu, 2001), and is suggested by Mwenifumbo et al. (2007) for the lower Manitou Falls Formation of the Athabasca Basin. The former abundance of detrital monazite is indicated by Th contents averaging 18

ppm (Quirt 1985) in the eastern Athabasca Basin associated with the lower Manitou Falls Formation, an average of 40 ppm around the Midwest Deposit (Ibrahim and Wu, 1985), abundant peaks between 20 and 50 ppm Th in borehole gamma-ray logs both near and away from deposits (geochemical analyses reach 730 ppm, Mwenifumbo and Bernius, 2007), and a broad Th anomaly on airborne gamma-ray maps (Carson et al., 2002a, b) that coincides with the entire distribution of the lower Manitou Falls Formation.

Could U have been carried adsorbed on Fe-Ti-oxides-hydroxides, hematite, altered zircon, and clay minerals (Macdonald, 1980; Hoeve and Quirt, 1984)? Such 'chemical sponges' have modern analogues in tropical soils and rivers. The lack of base metals in the unconformity-associated U ores is consistent with fluids leaching only the Athabasca Group because these metals are typically derived from feldspar that is absent from the Athabasca Group but still present in the basement. Inflowing surficial and ground waters could also have carried U (Hoeve and Sibbald, 1978; Macdonald, 1980; Hoeve and Quirt, 1984; Kyser et al., 2000). Unlike the basement, the clastic basin fill had very high permeability, huge volume, and abundant surface area on clastic grains of all types for chemical reaction. The degree of alteration of the Athabasca Basin overall (Kyser et al., 2000), and the infiltration of Cretaceous oil through the Fair Point Formation (Wilson et al., 2007) demonstrate that pervasive fluid flow affected a vast volume of the Athabasca Group.

#### Transport of Uranium

When, under what conditions, and how far was the U transported from its basement and/or detrital sedimentary sources? The timing of U movement in diagenetic fluids is illuminated by its content in phosphate minerals. Early diagenetic xenotime that overgrows zircon in the Athabasca Group contains virtually no U. Later diagenetic fluorapatite contains U in both Athabasca (Rainbird et al., 2003b) and Thelon (Miller, 1983) basins, yet predates the oldest uraninite ages. The apparently latest diagenetic AP minerals also lack U (Mwenifumbo and Bernius, 2007) but are interpreted to be about the same age as the ore deposits, recording conditions when U was released from precursor(s) and carried in solution elsewhere, presumably toward the ore deposits. Geochronology of U deposits and stratigraphy constrains ore formation to a period after deposition of sequence 3 to before deposition of sequence 4 of the Athabasca Group.

Geochemical factors constraining the development, movement, and mineral chemical changes accompanying fluids in sedimentary basins have been treated extensively in the literature (e.g. Hoeve and Quirt, 1984; Hiatt and Kyser, 2000; Kyser et al., 2000; Hiatt et al., 2003; Polito et al., 2004, 2005) for the three best-known Paleo- and Mesoproterozoic sedimentary basins that host unconformity-associated U deposits: Athabasca, Thelon, and Kombolgie (Fig. 6). The histories of fluid movement in these basins involved multiple low- to high-temperature events over hundreds of millions of years. Low-temperature uraniferous fluids are still in circulation. Diagenetic contrasts between the Athabasca, Thelon, and Kombolgie basins (Fig. 6) resulted in different mineral parageneses that record different equilibrium fluids (Kyser et al., 2000). However, in all three

cases, oxidized ( $fO_2 > -45$ , in the hematite field), saline (chlorinity up to 6 molal) basinal brines transported the U (Ruzicka, 1996a; Cuney et al., 2003). High  $fO_2$  is based on the lack of organic matter and the pervasive hematite in these basins.

The passage of later diagenetic reducing fluids is recorded by the bleached zone that invades the red regolith and red mudstone in the Read Formation, the drab grey and tan mudstone of the Manitou Falls Formation, and the presence of hydrocarbon and bitumen surrounding uraninite in many of the ore deposits. Proposed origins of the hydrocarbons range from the migration of hydrocarbons through the basin at least twice (Wilson et al., 2007) to abiotic synthesis (e.g. McCready et al., 1999; Sangély et al., 2003).

Acidity was controlled by the kaolinite-illite buffer to a pH of about 4.5 at 200°C (Cuney et al., 2003). Feldspar was either lacking or altered during diagenesis to form the sparse regional illite in the quartzarenite. Early diagenetic brines preserved as inclusions in quartz overgrowths on detrital quartz grains are NaCl-rich and inferred by Cuney et al. (2003) to have been derived from evaporitic layers that once existed in upper strata of the basin. Derome et al. (2002, 2003a, b) have determined that the brines trapped later in pervasively silicified zones and drusy quartz, close to the mineralized zones, became enriched in Ca, and inferred this to have resulted from their earlier interaction with Ca-rich basement rocks. High Ca in the mineralizing fluid is of major importance for accessory mineral alteration and for U mobilization from basement source rocks as shown by (i) incongruent dissolution of monazite with U-P-LREE leaching and new formation of a Th-U silicate with lower Th/U ratios, (ii) new formation of U-poor Ca-Sr-REE hydrated Al-phosphates, and (iii) Ca-REE-U-Al-P enrichment of zircon altered zones (Hecht and Cuney, 2000, 2003; Cuney et al., 2003). On the other hand, removal of significant Ca from basement rocks should have caused albitization, which has not been observed. Derome et al. (2003a) preferred a shallow source, above the preserved Athabasca Group, given the inferred 140°C temperature of the calcic fluid inclusions. Condition (i) above would have applied to alteration of detrital monazite in the Manitou Falls Formation and the Franceville Basin.

The effect of fluid compositions on U-solubility has not been quantified experimentally (Cuney et al., 2003). Uranium solubilities of 30 ppm were calculated by Raffensperger and Garven (1995) for five-molal Na-Ca-Cl solutions at 200°C for a  $fO_2$  of -20, well within the hematite field, and the concentrations of other possible strong U-ligands (e.g. F and P) is only limited by the solubility product of fluorite and apatite (Cuney et al., 2003).

Temperatures during primary mineralization are interpreted in various ways. Pagel et al. (1980), Kyser et al. (2000), and Cuney et al. (2003) interpreted that ore was deposited during peak diagenesis at 180 to 250°C, suggesting a geothermal gradient in the order of 35°C/km. Ramaekers (2004) suggested that either the geothermal gradient beneath the Athabasca Basin was anomalously high (40-50°C/km for a 5 km thick basin-fill) or that the basin-fill was much thicker before erosion. In contrast, fluid inclusion studies by Derome et al. (2003a) indicate that temperature

and pressure close to the unconformity decreased from the 'early diagenetic' 160 to 220°C and 1 to 1.25 kbar respectively from Rabbit Lake and Carswell deposits to the mineralization stage of 140 to 160°C and 0.6 kbar. Derome et al. (2003b) found that a late, low-saline, CH<sub>4</sub>-bearing, higher temperature fluid (200°C) was derived from the basement, and was commonly mixed with basinal NaCl brines in the Kombolgie Basin but is rarely recorded by fluid inclusions in the Athabasca Basin. Other fluid inclusion results have been obtained from widely distributed unconformity-associated U deposits such as in the Kombolgie Basin of Australia (Derome et al., 2003b) and Shea Creek (Derome et al., 2002), and work is in progress on fluid inclusions and other micro-analytical techniques for samples from Rabbit Lake and McArthur River (M. Cuney, pers. comm., 2005).

Uranium sourcing and transport were essentially independent of the local stratigraphy above the deposits, as shown by differences between the basal Read Formation and the overlying lower Manitou Falls Formation in the eastern Athabasca Basin. Whereas both formations are conglomeratic, the Read Formation has much lower Th, abundant red mudstone with desiccation cracks (both lacking in Manitou Falls Formation), few black laminae (locally abundant in Manitou Falls Formation), and this is the unit that directly overlies the deepest red regolith. Such attributes of the Read Formation record highly oxidizing conditions and subaerial exposure before and during sedimentation. The unconformably overlying Manitou Falls Formation records little evidence of oxidizing conditions or subaerial exposure during sedimentation, and its lower two members have high Th contents. Mwenifumbo et al. (2007) and Yeo et al. (2007) infer from this that the amount of monazite and other labile minerals that contributed to primary sedimentation was small in the Read Formation. Given that the Read and Manitou Falls formations have very similar diagenetic histories, it is highly unlikely that Th was selectively removed from the Read and upper Manitou Falls formations during diagenesis and/or hydrothermal alteration, or Th selectively added to the lower Manitou Falls Formation. The amount of U present as U-oxide in the original sediments of the Read Formation must have been very low owing to its lack of organic matter and inferred highly oxidizing conditions (Cuney et al., 2003). Thus the Read Formation is not considered to have been a good source of U, in contrast to the Fair Point and lower Manitou Falls formations. Large pathways for mineralizing fluid flow are necessitated by these compositional differences and highlighted by the fact that the barren Read Formation has a limited distribution but significant deposits are independent of that. The Read Formation directly overlies the McArthur River ore pods, but is absent over the Cigar Lake deposit.

### Focus of Uranium Deposition

Aquifers along the unconformity, brittle reactivated faults (including seismic pumping), crosscutting local structures and alteration (e.g. silicification, clay minerals, and dissolution) were the main controls on fluid flow at the deposit sites. The graphitic metapelitic gneiss units are not only conductive targets and the sites of reactivated faults, but are also widely regarded as a key source of reductant in geochemical process models for unconformity-associated U, albeit with-

out consensus (e.g. Hoeve and Sibbald, 1978; McCreedy et al., 1999; Wilson et al., 2007). Graphitic units may also have focused U precipitation from hydrothermal fluids by conducting deep heat sources to drive convection (Hoeve and Qirt, 1984) or by serving as anodes of natural electrical systems. Significant U deposits can form in the absence of graphitic units (e.g. Kiggavik, Fuchs and Hilger, 1989; and some of the deposits at Cluff Lake), however these are in the minority and it is not known whether super high-grade deposits, such as McArthur River, can form without graphite.

Intersections of reactivated basement shear zones with offsets of the unconformity and intersections between different fault arrays enhanced fluid flow to focus U deposition, and now guide mine-scale exploration and development. Studies relating fault intersections, inferred fluid flow, and ore locations include Baudemont and Paquet (1996) at McClean Lake; Baudemont and Federovich (1996) at Cluff Lake; and Rhys (2002), D. Brisbin (oral presentation, 32nd International Geological Congress, Florence, Italy, August 20-28, 2004), R. LeMaitre and C. Belyk (oral presentation, Saskatchewan Geological Survey, Open House 2004, November 30, 2004, Saskatoon, Saskatchewan), and D. Thomas (oral presentation, Saskatchewan Geological Society, Uranium Short Course, November 29, 2005, Saskatoon, Saskatchewan) regarding new discoveries at Eagle Point. Similar work during active mining at Sue C Pit, McClean Lake (Tourigny et al., 2002, 2007) showed that the south-raking geometry of elongate ore lenses and pods, together with structural elements in the enclosing shear zone, can predict the overall south-raking geometry of the deposit. They have suggested that en echelon arrays of uraninite veins at Sue C Pit may represent mineralized hybrid extensional shear fractures.

The fault intersections related to at least two ore deposits coincide with paleovalleys that are reflected in thickness, facies changes, and paleocurrents in the Read and lower Manitou Falls formations (Harvey and Bethune, 2007; Long, 2007). Recognition of such valleys provides an additional focus for exploration. Such paleovalleys, the linear geometry of their river channels, and basement ridges may also have influenced fluid flow related to ore formation (Collier and Yeo, 2001). Faults and fracture systems are still open in mine districts and form important present-day aquifers that must be accounted for in mine development and environmental monitoring of groundwater. The fracture systems affect present day thermal conductivity (Mwenifumbo et al., 2004) and host recent limonitic alteration products. Thermal anomalies caused by the conductive properties of graphitic metapelite would have focused upward fluid flow in their vicinities.

### **Knowledge Gaps**

Why would such high-grade, large-tonnage U deposits be found only at the basal unconformities of shallow, late Paleo- to Mesoproterozoic conglomeratic sandstone basins? And why does the Athabasca Basin host deposits that are one or two orders of magnitude larger than in similar basins elsewhere? Further mineralogical and litho-geochemical analysis is needed to test, expand, and focus the three-basin comparison of Kyser et al. (2000), which assigns higher overall U potential to basins with more intense alteration histories. An

alternative perspective should also be pursued, i.e., that these different basins may have similar U potential but require different exploration paradigms adapted to their different alteration characteristics.

Whereas it might seem that exploration is at a mature stage in the Athabasca Basin, this single basin is larger than some Canadian provinces and many countries of the world, and only a small part of it has been touched by intensive exploration. Almost every year for the past 30 years, a significant discovery or advancement has been made there. Is the entire basal unconformity surface prospective where it intersects favourable basement domains and reactivated graphitic shear zones? Will additional knowledge provide tools to expand production from the existing clusters of deposits and significant prospects?

#### *Origins of Intracontinental Proterozoic Basins*

The triggers and drivers for the development of intracontinental Proterozoic basins and their fluid histories have long been considered enigmatic (Ross 2000). It is becoming clear that the Athabasca Basin developed by late-stage transpressive tectonic processes (Ramaekers, 2004; Ramaekers et al., 2007). Ruzicka (1996a) used terms such as “rapid subsidences” and “rifting” to describe events that triggered hydrologic systems, however such events were neither as “rapid” nor “rifting” as dramatic as in continental rift basins or strike-slip basins – these intra-continental events involved subtle, gentle subsidence, uplift just sufficient to generate cobble and pebble conglomerate, and the rifting led to accommodation space for just slivers of sediment accumulation compared to passive margin basins. Nonetheless, it was indeed tectonism, subtle tilts in the basin floor and reactivation of bounding faults that must have both driven and focused hydrothermal circulation to form the unconformity ore deposits. Much work remains to document the relationships between the different orders of fault systems and their orientations, to determine which faults are most prospective and when they focused ore-forming fluids. The generally accepted protracted fluid history in the Athabasca Basin, the wide range in uraninite ages, and the older regional phosphatic alteration of the Athabasca Basin challenge researchers to tackle regional background samples (e.g. Pagel, 1975) to help place the various alteration and putative ore forming fluid events into a basin-development framework.

#### *Paleoweathering*

Is the red-green basement alteration below the unconformity due to post-sedimentary alteration (Cuney, 2003) or a superimposition of such alteration on a primary paleoweathering profile (McDonald, 1980; Hoeve and Quirt, 1984)? Textural evidence of paleosol should include geopetal features and intraclasts of paleosol with red-green zonation preserved in the basal conglomerate. Paleosols are present within the lower Thelon and Manitou Falls formations (e.g. Hiatt et al., 2003; Hiatt and Kyser, 2007), and diaspore has been recognized within the regolith (Hoeve and Quirt, 1984). However in mineralized areas where most data have been collected, the ‘white clay’ (illite) alteration, where present along the unconformity, extends up into the basal conglomeratic sandstone units and obliterates most paleosol features

that might have been preserved in lithic fragments. Well constrained fieldwork and geochemistry comparing the unconformity assemblage proximal and distal to ore would help answer these questions and develop additional exploration vectors.

#### *Mineralogical Anomalies*

Regional illite appears to be developed only in certain corridors where basement structures crosscut the basin, such as the McArthur River - Key Lake corridor (Fig. 7) and at Shea Creek (Rippert et al., 2000). Where no basement structure is observed, as in the Erica 1 or Rumpel Lake drillholes, most of the primary kaolinite, or dickite - its regional alteration product - are preserved, and illite is poorly developed.

Regional illite+kaolinite assemblages represent terrestrial strata (most of the Athabasca Group) whereas illite+chlorite (suduïte), with minor expandable layers in illite, represent marine strata (Wolverine Point and Douglas formations) (Hoeve and Quirt (1984, p. 38-44). Potassium was conserved in the regional diagenetic process and would not have been moved in quantity to form the illite-dominated deposit-related alteration halos. Potassium for illite, as well as Mg and Fe for chlorite in these alteration halos, must have come from basement-derived fluids. Separate linear tourmaline and chlorite alteration zones (see Fig. 7) suggest discrete basement sources for B versus Mg+Fe. How are these related to U potential?

How critical is the degree of quartz domination in the Athabasca Basin compared to correlative basins? A corollary is the importance of possible detrital U carriers such as clay, iron hydroxide, and heavy minerals such as monazite. Do these result from primary sedimentary, diagenetic-alteration-fluid flow, and/or incomplete sampling histories? Do such differences mean different overall U potential or just different exploration strategies?

#### *Geochronology*

U-Pb ages on U oxides have errors of several to tens of millions of years, attributed to continuous or/and episodic diffusion of radiogenic Pb out of U oxides (Cuney et al., 2003). It is unknown whether the common ore ages of about 1350, 1000, and 300 Ma are all reset from older primary ages of 1600 to 1500 Ma that have been determined recently on the McArthur, Cigar Lake, and Sue deposits (e.g. Cumming and Krstic, 1992; Fayek et al., 2002a, b; Alexandre et al., 2003).

Calculation of the time necessary to form these massive uraninite orebodies requires numerous parameters, most of which are poorly constrained. Assuming that the mineralizing fluid contained 5 to 10 ppm U, percolated at rates of approximately 0.1 m/year, and had a volume of several tens of cubic kilometres, the formation of Cigar Lake would have required a few million years. In contrast, the Sue C and McArthur River deposits developed as lenses and pods within low-pressure dilatant jogs of transpressive faults by active processes such as seismic pumping (Sibson, 2001; Tourigny et al., 2007), also known as fault valve behaviour (Nguyen et al., 1998). The duration of ore formation was constrained by the duration of fault activity and the effectiveness of U transport.



Typical second-order basin-filling sequences are thought to require 20 to 45 million years for deposition (Krapez 1996); nevertheless, the duration of Precambrian examples is difficult to determine with such precision. The 1740 to 1760 Ma maximum age of the Athabasca Group and the 1644 Ma age of the tuff in sequence 3 allow about 100 Ma for deposition and development of the unconformities at the tops of sequences 1 and 2, and basin-wide phosphate diagenesis. Deposition of upper sequence 3, erosion and deposition of lower sequence 4 (Douglas Formation), pre- and post-ore alteration and mineralization took place during the next 100 Ma, with repeated U remobilization taking place over hundreds of millenia. Will it be possible to temporally correlate these sequences and mineralization events with those of the Thelon, Hornby Bay, and Elu basins? Will the older uraninite ages of ca. 1680 Ma for the Kombolgie Basin (Polito et al., 2005) be discovered in Canadian basins?

### *Fluid Flow*

Is it possible to quantify what different fluid-flow events took place, what were their paths, and which events were responsible for ore deposition over the remarkable 100s of millions of years of alteration in Proterozoic basins? Unpublished modern fluid-flow modeling (Ord, 2003) has been built on a rich base of ideas and data (e.g. Hoeve and Quirt, 1984; Sibbald, 1985; Hoeve and Quirt, 1987; Wilde and Wall, 1987; Quirt, 1989; Bruneton, 1993; Fayek and Kyser, 1997; Kyser et al., 2000; Jefferson et al., 2001; Cuney et al., 2003; Kister et al., 2003; Polito et al., 2004; Ramaekers, 2004; Hiatt and Kyser, 2007; Ramaekers et al., 2007).

How were fluid flows balanced during downward (ingress) and upward and outward (egress) to and from basement fault zones? Was it rectilinear with long horizontal flow paths? Both modern and ancient flow paths reach more than 1000 m depth within the basement, are significant along the basal unconformity and in the regolith, and extend throughout the full preserved thickness of the Athabasca Group up to the Wolverine Point Formation aquitard. The Douglas Formation mudstone aquitard may also have constrained convection. Modern fluid flow around Cigar Lake and McArthur River is along subhorizontal aquifers parallel to bedding with shorter flow paths in near-vertical to 45° fault zones (e.g. Cramer and Smellie, 1992). Ore deposits formed where stable ingress or egress zones fostered prolonged mixing of U-bearing oxidized basinal fluids with reducing basement fluids. An ongoing challenge is to distinguish sites of ore-related focused flow from sites of diffuse fluid flow.

### *Fluid Chemistry: Causes of Quartz Dissolution and Uranium Precipitation*

A basement-derived, reduced fluid was proposed by Hoeve and Sibbald (1978) in the mixing model to explain U deposition, the source of Mg (dravite and sudoïte alteration), B (dravite) in the sandstone, and Ni, Co, Cu, Zn, and Au in the polymetallic deposits. What reducing agents could have come from graphitic metapelite? Hydrocarbons (McCready et al., 1999; Annesley et al., 2001) are unlikely because pure graphite and water do not react below 400°C, such a reaction is thermodynamically unlikely, and bitumen paragenesis is overwhelmingly post-uraninite (Leventhal et al., 1987; Wilson et al., 2007). Could disorganized graphite grains

from the basement have reacted with Na-Ca-Cl brines at low temperature to form the CH<sub>4</sub> and N<sub>2</sub> that have been found in fluid inclusions (e.g. Hoeve and Quirt, 1987; Landais et al., 1993; Cuney et al., 2003; Sangély et al., 2003)?

Some deposits are not associated with graphitic metapelite, e.g., the Kiggavik and several Cluff Lake deposits are not associated with graphitic basement rocks. Consideration should be given to the sulphide and mafic mineralogy of basement rock, the activity of Fe<sup>2+</sup> (Cramer, 1986; Wilde and Wall, 1987; Quirt, 1989 and earlier workers cited therein; Wilde et al., 1989) and decrease in pH of an alkaline fluid, with iron being the electron acceptor to convert U<sup>6+</sup> to U<sup>4+</sup> (S. Romberger, Uranium short course, Cordilleran Roundup Vancouver, British Columbia, January 20, 2006), heat flow (Hoeve and Quirt, 1984), and electrochemical potential.

The basement fluid must also have been alkaline and under saturated in silica to cause the ubiquitous quartz dissolution above ore deposits. Slightly warmer basement fluid in equilibrium with rock such as pelitic gneiss would, upon introduction into the sandstone, be under saturated in silica (Hoeve and Quirt, 1984, 1987). Is such a fluid capable of destroying zircon as witnessed at Cigar Lake?

### *Triggers for Uranium Deposition*

For every unconformity-hosted deposit underlying an egress-type alteration zone, how many corresponding basement-hosted deposits might exist with ingress-type alteration? Basement-hosted, monomineralic deposits are difficult to find but are attractive targets because they are suitable for open pit mining if close to surface and more importantly, because they are hosted in relatively competent basement rocks, there is much less need for freeze-wall technology to control water, even if deep; in addition, they are metallurgically attractive. Some deposits may remain to be discovered in areas of shallow Athabasca Group cover. They are unlikely to be preserved more than a few kilometres outside the eastern Athabasca Basin given the depth of erosion (Harper and Yeo, 2005), however, extensive areas of thin regolith around the Thelon Basin remain prospective, with the Kiggavik deposit being one example. Nisto, a small past producer northeast of Black Lake (Fig. 4; Appendix 1; Macdonald et al., 2000), is one possible example outside the Athabasca Basin.

If both ingress- and egress-type fluid flow developed along reactivated basement fault systems, and if hydrothermal convection is integral to the genesis of unconformity-associated U deposits, each fault system that generated a basal sediment- and/or regolith-hosted ore deposit had the potential to generate basement-hosted deposits. Geophysical, geochemical, and mineralogical tools are being continually improved and reapplied in this regard. Reevaluation of historical exploration, particularly drilling, is being undertaken with this difficult model in mind. Process models involving seismic pumping may be tested by further structural analysis, and this may provide structural geological tools to locate sites of ingress.

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