UNCONFORMITY ASSOCIATED URANIUM DEPOSITS
DRAFT

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Definition

Unconformity-associated uranium (U) deposits comprise massive pods, veins and/or disseminations of uraninite spatially associated with unconformities between Proterozoic siliciclastic basins and metamorphic basement. The siliciclastic basins (Figure 1) are relatively flat-lying, un-metamorphosed, late Paleoproterozoic to Mesoproterozoic, fluvial red-bed strata. The underlying basement rocks comprise tectonically interleaved Paleoproterozoic metasedimentary and Archean to Proterozoic granitoid rocks. Uranium as uraninite (commonly in the form of pitchblende) is the sole commodity in the monometallic sub-type and principle commodity in the polymetallic sub-type that includes variable amounts of Ni, Co, As and traces of Au, Pt, Cu and other elements. Some deposits include both sub-types and transitional types, with the monometallic tending to be basement-hosted, and the polymetallic generally hosted by basal siliciclastic strata and paleo-weathered basement at the unconformity.

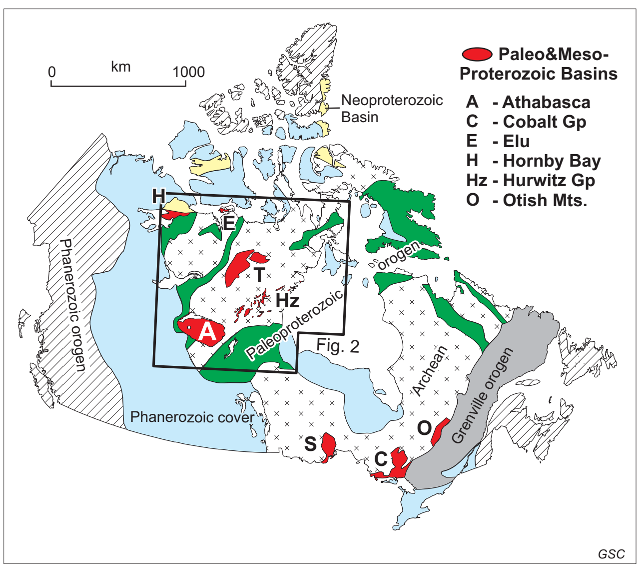

FIG. 1. Canadian Paleo-Mesoproterozoic basins (black) within the Canadian Shield have potential for or contain known unconformity-associated U deposits. Selected Meso- to Neoproterozoic basins are shown by dotted pattern.

Grade, Tonnage and Value Statistics

Global Unconformity-Associated And Other Uranium Resources

Global context and consideration of all U deposit types are important for those engaged in production, exploration and outreach in the Canadian nuclear energy industry, because social-political aspects of the nuclear energy industry strongly affect its viability around the world (Canadian Nuclear Association, www.cna.ca; Uranium Info Centre Ltd., www.uic.com.au). World U resources are contained in some fourteen different deposit types, with the major types in decreasing order of world resources as follows: Mesoproterozoic unconformity associated (>33% in Australia and Canada), the one giant Olympic Dam Mesoproterozoic breccia complex deposit in Australia (>31%), sandstone hosted (18%, mostly in the USA, Kazhakstan and Niger), surficial deposits (4% mainly in Australia), large tonnage but low grade resources in early Paleoproterozoic conglomeratic deposits, and small percentages in volcanic, metasomatic, metamorphic, granite-hosted and vein-type deposits (World Uranium Mining, 2004).

Uranium resource data for Canadian and comparative Australian Mesoproterozoic unconformity-associated deposits are compiled in a digital database by Gandhi (2005). The starting point for most of the Saskatchewan deposits is the Geological Atlas of Saskatchewan (Saskatschewan Industry and Resources, CD-ROM, version 5, by W. Slimmon) that includes a digital database. Table 1 and Figure 2 here summarize individual grades and tonnages of 42 Canadian and Australian deposits from the digital database, and Table 2 provides totals for the Athabasca and Thelon basins (Fig. 3). The Hornby Bay and Elu basins are less well explored and no unconformity-associated resources have been outlined in them, although the Hornby Bay Group hosts one sandstone-type deposit with drill-indicated

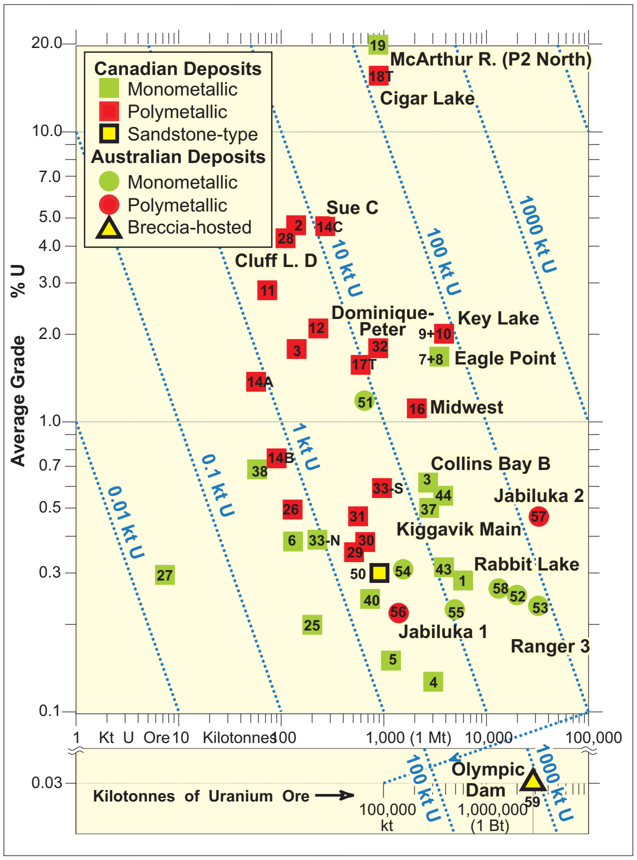

FIG. 2. Grade/tonnage plot of Mesoproterozoic U deposits of the unconformity-related type and selected other types in Canada and Australia (Table 1; after Ruzicka, 1996a and Gandhi 1995). Numbers correspond to Table 1. Selected deposits are named.

resources. Eight Australian Mesoproterozoic unconformity-associated deposits are included in Table 1 and summarized in Table 2 for comparison with Canadian deposits. Also included is the exceptional Olympic Dam deposit (see production details below) that has complex mineralogy and is genetically related to the host Gawler Range continental volcano-plutonic complex.

Uranium resources (Table 1) are listed in metric units, according to IAEA practice (Anonymous, 2003a). The chemical formula "U_3O_8" is commonly used instead of "U metal" by industry and some government sources, because the product from mining and milling is "Yellow Cake", ammonium diuranate. It is later calcined to produce U oxide, a dark grey-green powder that assays slightly higher than 99 % U_3O_8. Early data on resources and production were in imperial units. The practice is still followed, although it often leads to a mixture of units of the two systems e.g., 12.3 million lbs of U_3O_8 in 229.3 thousand tonnes of ore (viz., metric tonnes of ore) grading 2.43 % U_3O_8. The factors for conversion to the metric units are: 1 % U_3O_8 = 0.848 % U; 1 lb U_3O_8 = 0.3846 kg U; 1 short ton = 0.90718 tonne; and 1 lb U_3O_8 per short ton = 0.4240 kg/tonne. In some cases grade is expressed as kg/t, which should not be confused with 'per cent' because 1 kg/t is 0.1 %.

Resource values presented in the tables are 'global' or 'mineable' for the deposits that are unmined or partially mined. For the deposits that are mined out the values represent the actual production. Geoscience data for deposits in the Athabasca and Thelon basins (Table 1) are in a digital database (Gandhi, 2005). No attempt is made here to classify the resources in different categories using economic parameters, although this was an important task for the 'Uranium Resource Assessment Group' at the GSC of which Gandhi was a member for 20 years. Cut-off grades, if available in the published data, are noted in the database. In most of the unconformity-related deposits the ore-waste boundary is commonly sharp or narrow, especially for the high-grade deposits. **Caveat:** *Reserves/Resources reported here and by Gandhi (2005) are historical values that are not compliant with Canadian legislation set out in National Instrument 43-101 (Canadian Securities Administrators, 2001a, b). Hence the original data source should be cited when re-reporting them.*

Over the last decade data have changed for some well-known deposits, and some new deposits have come to light. A logarithmic plot (Figure 2) clearly illustrates the current relative importance of various world-class deposits, with the following highlights summarized from the digital compilation (Table 1; Gandhi, 2005).

- Unconformity-related deposits of the Athabasca basin are the world's largest storehouse of high-grade U resources. The most spectacular grades and tonnages are those of the Cigar Lake and McArthur River deposits, which are 15 and 22.28 % U respectively and contain 131,400 and 192,085 tonnes U, respectively (Fig. 2). The average grade for some 30 unconformity deposits in the Athabasca basin, including these two high-grade examples, drops to

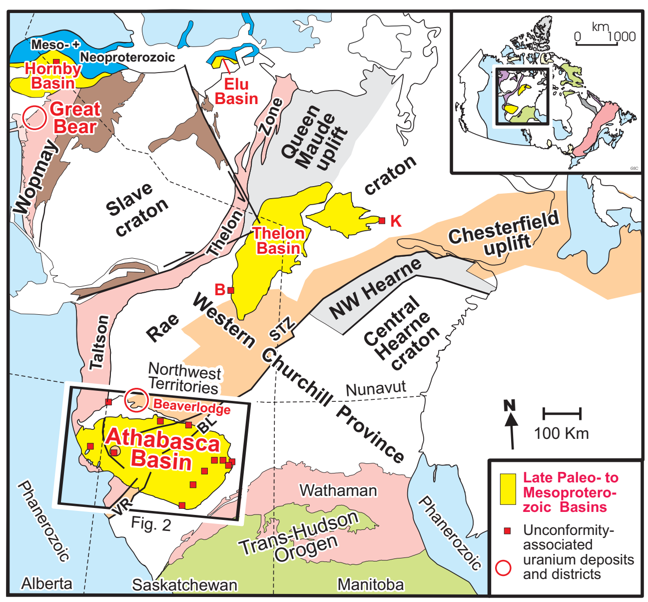

FIG. 3. Relationships of the Athabasca Basin to major tectonic elements of the northwestern Canadian Shield, after Thomas et al. (2000) and Western Churchill Metallogeny Project team. Unconformity -associated prospects of the Thelon Basin are Boomerang Lake (B) and Kiggavik (K). Hornby Basin sandstone-hosted prospect is PEC-YUK. Deposits and prospects are listed in Table 1, their grades and tonnages plotted in Figure 2, Athabasca Basin deposits are located in more detail on Figure 4 (keyed to oblique rectangle), and variations in mineralization and alteration style shown in figures 5 and 6. The classic vein-type deposits of the Beaverlodge and possibly Great Bear Lake U districts are here considered as exhumed roots of unconformity-associated deposits.

Table 1. Resources of unconformity-related and other selected uranium deposits plotted in Fig. 2. Summarized from compilation by Gandhi (2005).

No.*	Basin	Deposit / Prospect	Latitude	Longitude	Discovery	Ore (kt)	U Grade (%)	Tonnes U	Status
1	Athabasca (Eastern)	Rabbit Lake mine	58° 11' 53" N	103° 42' 40" W	1968	5840	0.27	15,769	mined out
2	Athabasca (Eastern)	Collins Bay A	58° 17' 02" N	103° 37' 37" W	1971	134.6	4.83	6,500	mined out
3B	Athabasca (Eastern)	Collins Bay B	58° 15' 50" N	103° 38' 49" W	1977	2582	0.61	15,750	mined out
3D	Athabasca (Eastern)	Collins Bay D	58° 16' 55" N	103° 38' 25" W	1979	129.5	1.66	2,150	mined out
4	Athabasca (Eastern)	Raven	58° 09' 05" N	103° 45' 47" W	1972	3042	0.12	3,650	
5	Athabasca (Eastern)	Horseshoe	58° 09' 28" N	103° 44' 17" W	1974	1107	0.14	1,550	
6	Athabasca (Eastern)	West Bear	57° 52' 32" N	104° 03' 51" W	1977	121.6	0.37	450	
7	Athabasca (Eastern)	Eagle Point: South zone	58° 17' 44" N	103° 36' 14" W	1980	2050	1.39	28,850	depleted
8	Athabasca (Eastern)	Eagle Point : North, 01, 02, 03	58° 17' 44" N	103° 36' 14" W	1980	1267	1.76	22,300	depleted
7,8	Athabasca (Eastern)	Eagle Point: total zones	58° 17' 44" N	103° 36' 14" W	1980	3317	(1.542)	(51,150)	depleted
9	Athabasca (Eastern)	Gaertner	57° 12' 11" N	105° 39' 58" W	1975	1345	1.71	23,000	mined out
10D	Athabasca (Eastern)	Deilmann	57° 12' 35" N	105° 37' 59" W	1975	2242	2.11	47,300	mined out
10C	Athabasca (Eastern)	Cobble 0re (in glacial drift)	57° 12' 23" N	105° 38' 59" W	1975	~150	~2	~3,000	mined out
2371	Athabasca (Eastern)	*P Patch (Boundary Lake east of P grid, DDH P14&15)*	*57° 12' 39" N*	*105° 34' 13" W*	*1999*	*na*	*na*	*na*	
9,10	Athabasca (Eastern)	(Total: Key Lake deposits)	57° 12' 23" N	105° 38' 59" W	*1975*	3695	(2)	(73,900)	mined out
11	Athabasca (Eastern)	JEB deposit	58° 20' 57" N	103° 50' 10" W	1982	71.9	2.79	2,003	mined out
12	Athabasca (Eastern)	McClean N, SE and SW pods	58° 15' 47" N	103° 50' 49" W	1978-'79	71.2	4.58	3,256	
13	Athabasca (Eastern)	(McClean S = SE + SW pods)	58° 15' 26" N	103° 50° 49" W	1978-'79	158.2	0.933	1,475	
14A	Athabasca (Eastern)	Sue A zone	58° 15' 42" N	103° 48' 11" W	1988-'89	55	1.26	692	
14B	Athabasca (Eastern)	Sue B zone	58° 15' 54" N	103° 48' 06" W	1988-'89	90.1	0.73	654	
14C	Athabasca (Eastern)	Sue C zone	58° 15' 36" N	103° 48' 11" W	1988-'89	249.9	4.5	11,247	mined out
12-14	Athabasca (Eastern)	(Total: McClean L. deposits)	58° 15' 49" N	103° 48' 11" W	*1980-'89*	696.3	(2.78)	(19,327)	
15	Athabasca (Eastern)	*Caribou (McClean new zone)*	58° 15' 50" N	103° 49' 20" W	*2002*	*na*	*na*	*na*	
16	Athabasca (Eastern)	Midwest (+ 3 U showings)	58° 18' 11" N	104° 05' 10" W	1978	2033	1.06	21,550	
17ST	Athabasca (Eastern)	Dawn Lake 11, 11A, 11B, 14	58° 19' 35" N	103° 55' 55" W	1978				
17T	Athabasca (Eastern)	*(Total: Dawn Lake zones)*	58° 20' 12" N	103° 55' 41" W	1978	*685*	*(1.671)*	*(11,446)*	
18E	Athabasca (Eastern)	Cigar Lake: East zone	58° 04' 22" N	104° 32' 17" W	1981	557	15.4	85,766	
18W	Athabasca (Eastern)	: West zone	58° 04' 22" N	104° 32' 17" W	1981	318	14.35	45,620	
18T	Athabasca (Eastern)	(Total: East + West zones)	58° 04' 22" N	104° 32' 17" W	*1981*	875	(15.02)	(131,386)	
19	Athabasca (Eastern)	McArthur River / P2 North	57° 46' 07" N	105° 04' 55" W	1988	1017	22.28	192085	
20	Athabasca (Eastern)	*Paul Bay prospect*	*57° 50' 39" N*	*104° 52' 26" W*	*1989*	*na*	*na*	*na*	
21	Athabasca (Eastern)	Moore Lakes prospect	57° 27' 30" N	105° 09' 15" W	2000	*na*	*na*	*na*	
22	Athabasca (Eastern)	*La Rocque Lake prospect*	*58° 29' 45" N*	*104° 34' 45" W*	*2000*	*na*	*na*	*na*	
23	Athabasca (Eastern)	*Millenium prospect*	*57° 37' 30" N*	*104° 47' 00" W*	*2002*	840	2.304	19370	
24	Athabasca (North Rim)	Middle Lake prospect	59° 13' 13" N	105° 32' 35" W	1952	*na*	*na*	*na*	
25	Athabasca (North Rim)	Fond-du-Lac prospect	59° 18' 22" N	107° 10' 01" W	1967	200	0.2	400	
26	Athabasca (North Rim)	Maurice Bay prospect	59° 23' 07" N	109° 54' 00" W	1977	120	0.5	600	
27	Athabasca (North Rim)	Stewart Island prospect	59° 19' 47" N	108° 53' 49" W	1953	7.4	0.3	22	
1621	Athabasca (North Rim)	*Nisto deposit (past producer)*	*59° 12' 37" N*	*105° 26' 40" W*	*1948*	*na*	0.24	*na*	mined out
28	Athabasca (Western)	Cluff Lake D zone	58° 21' 52" N	109° 31' 01" W	1969	107.6	4.133	4,447	mined out
29	Athabasca (Western)	Cluff Lake N zone	58° 22' 23" N	109° 29' 20" W	1969	505	0.34	1,729	mined out
30	Athabasca (Western)	Claude	58° 23' 12" N	109° 32' 10" W	1971	640	0.367	2,349	mined out
31	Athabasca (Western)	Cluff Lake OP zone	58° 22' 05" N	109° 31' 01" W	1970	25.5	0.425	108	mined out
32	Athabasca (Western)	Dominique-Peter	58° 22' 37" N	109° 31' 41" W	1980	868.1	0.644	5,591	mined out
33N	Athabasca (Western)	Dominique-Janine: North mine	58° 22' 07" N	109° 32' 31" W	1980	230	0.38	874	mined out
33S	Athabasca (Western)	D-J South and West mines	58° 22' 05" N	109° 32' 31" W	1984	950	0.58	5,510	mined out
28-33	Athabasca (Western)	(Total: Cluff Lake deposits)	58° 22' 37" N	109° 31' 41" W	1969-84	3326.2	(0.62)	(20,608)	mined out
34	Athabasca (Western)	Shea Creek Main or Anne Z.	58° 14' 15" N	109° 30' 00" W	1992	*na*	*na*	*na*	
35	Athabasca (Western)	*Shea Creek Colette prospect*	*58° 14' 32" N*	*109° 31' 05" W*	*1992*	*na*	*na*	*na*	
36	Athabasca (Western)	Maybelle River prospect	58° 12' 15" N	110° 39' 31" W	1977	*na*	*na*	*na*	
37	Thelon (Eastern)	Kiggavik Main	64° 26' 30" N	97° 38' 50" W	1977	2425	0.492	11,931	
38	Thelon (Eastern)	Kiggavik Centre	64° 26' 38" N	97° 37' 55" W	1978	560	0.661	3,702	
39	Thelon (Eastern)	Kiggavik East	64° 26' 42" N	97° 37' 05" W	1978	990	0.05	495	
40	Thelon (Eastern)	*Bong Grid prospect*	*64° 24' 53" N*	*97° 42' 12" W*	*1986*	*717*	*0.23*	*1,650*	
41	Thelon (Eastern)	*Granite Grid prospect*	*64° 25' 43" N*	*97° 46' 57" W*	*1992*	*na*	*na*	*na*	
42	Thelon (Eastern)	*Sleek Grid prospect*	*64° 21' 42" N*	*97° 50' 12" W*	*1992*	*na*	*na*	*na*	
43	Thelon (Eastern)	End Grid	64° 20' 39" N	97° 51' 54" W	1987	3722	0.308	11,463	
44	Thelon (Eastern)	Andrew Lake	64° 20' 00" N	97° 53' 45" W	1988	3575	0.539	19,269	
45	Thelon (Eastern)	*Jane 2 Grid prospect*	*64° 18' 33" N*	*97° 56' 03" W*	*1993*	*na*	*na*	*na*	
46	*Thelon (Southwest)*	*Boomerang Lake prospect*	*62° 38' 25" N*	*104° 59' 50" W*	*1983*	*na*	*na*	*na*	
47	*Thelon (Northern)*	*Deep Rose (U-rich boulders)*	*65° 12' 00" N*	*98° 55' 00" W*	*1975*	*na*	*na*	*na*	
48	*Thelon (Northern)*	*Garry Lake (U-rich boulders)*	*65° 40' 00" N*	*100° 00' 00" W*	*1981*	*na*	*na*	*na*	
49	*Ellis River*	*Twin Lakes (U-rich boulders)*	*66° 35' 30" N*	*108° 00' 00" W*	*1975*	*na*	*na*	*na*	
50	Hornby Bay	PEC-YUK (sandstone-type)	67° 20' 00" N	116° 57' 00" W	1969	900	0.3	2,700	
51	Kombolgie	Nabarlek	12° 18' 40" S	133° 19' 12" E	1970	721	1.276	8,967	mined out
52	Kombolgie	Ranger 1	12° 40' 23" S	132° 55' 05" E	1969	19780	0.271	46,640	mined out
53	Kombolgie	Ranger 3	12° 40' 23" S	132° 55' 05" E	1969	30933	0.22	68,053	depleted
54	Kombolgie	Ranger 68	12° 31' 00" S	132° 52' 00" E	1970	1500	0.303	4,545	
55	Kombolgie	Koongarra 1 and 2	12° 52' 12" S	132° 50' 24" E	1970	2481	0.564	13,992	
56	Kombolgie	Jabiluka 1	12° 30' 07" S	132° 53' 45" E	1971	1300	0.212	2,883	
57	Kombolgie	Jabiluka 2	12° 30' 07" S	132° 53' 45" E	1971	31100	0.449	138,224	
58	Paterson	Kintyre	22° 45' 00" S	122° 15' 00" E	1985	12200	0.25	30,500	
59	Stuart Shelf	Olympic Dam (Breccia-hosted)	30° 26' 25" S	136° 53' 22" E	1975	2877610	0.03	863,283	depleted

kt: kilotonnes; na: not available (no resource estimate); *Numbers of Athabasca deposits are keyed to locations in Figure 4.

Deposits/prospects in italics are not fully entered in Gandhi (2005); the four-digit numbers are from Saskatchewan SMDI files and are not in the database.

Summarized from database compilation by Gandhi (2005), whose sources include Anonymous (2003, 2005), McKay and Miezitis (2001); Thomas et al. (2000); Yeo and Jiricka (2002); and corporate annual reports, web sites and personal communications from COGEMA Resources Inc., Cameco Corp., PNC Exploration Canada Ltd., JNR Resources Ltd. and Western Mining Corp.

Table 2. Summary of U Resources in major Mesoproterozoic districts of northwestern Canada (red) and Australia.

District	Ore (Kt)[1]	% U[2]	Tonnes U
Athabasca Basin	28,810	1.922	553,778
Beaverlodge District[3]	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[4]	2,877,610	0.03	863,283

Data from Table 1: [1]Includes past production; [2]Calculated from Kt ore and Tonnes U, rounded to significant digits;[3]Past production from two classic vein-type (Eldorado and Lorado Mills) and one episyenite-type (Gunnar) deposits; [4]In Gawler Range volcano-plutonic complex. Olympic Dam has nothing to do with unconformity-associated U deposits, but is included here for comparison because it is such a vast resource of U, and it is approximately the same age.

1.97 % U, still four times the average grade (0.44 % U) of Australian unconformity-associated deposits. The Athabasca average is also more than an order of magnitude greater than that of classic vein-type deposits of the Beaverlodge district at the northwest margin of the basin (Smith, 1985, p. 99).

- The area of the Athabasca basin is more than 85 000 sq km, yet 96 % of the known U resources of unconformity-related type in it are concentrated in less than 20 % of the area along the eastern margin of the basin. This presents a metallogenic challenge for any reasonable assessment of the potential of the remaining 80 % of the basin.
- The Thelon Basin is nearly equal in area to the Athabasca basin and is geologically very similar (Table 3). The known U resources in it are 9 % of those in the Athabasca basin. Furthermore these are concentrated in the Kiggavik Trend, which is located in an area less than 500 sq km. The average grade of Thelon deposits is modest relative to those of Athabasca deposits, but is comparable to that of the Kombolgie Basin.
- Uranium metal resources of the Kombolgie Basin (or Carpentarian Basin with basal Kombolgie Formation sandstone) are slightly more than 50% of those in the Athabasca Basin Their ore tonnage is greater but their average grade is relatively low. These deposits are confined to an area of about 7500 km², known as the Alligator Rivers U field. This field shows geological similarities with the eastern Athabasca basin.
- Kintyre deposit is also a large tonnage and low-grade resource comparable with the deposits in the Kombolgie basin. Furthermore much of its mineralization is basement-hosted, as is the case for the deposits in the Alligator River field.
- Other U deposit types in the world contain very large resources, but are economic due to factors other than grade. A few of these are included as context, to help understand the phenomenal grades of unconformity-associated deposits. The Olympic Dam deposit is the largest U resource in the world but it is also the lowest in grade and essentially a one-of-a-kind Cu deposit from which U is won as a by-product along with REE, gold, silver and many other commodities. The present annual production capacity of this underground deposit is limited to 4,500 t U from ore grading close to 1.6% Cu, 0.06 % U, 0.6 g/t Au and 6 g/t Ag. Plans are to increase it to the level of 6,500 t U. This tonnage must be viewed in the context of the world production of 36,112 t U in year 2000, and the demand of 64, 014 t U for the world's 438 commercial nuclear reactors (NEA-IAEA, 2001). Other very large but low-grade U resources of the world include the volcanogenic Streltsovka caldera in Russia with 250,000 t U, sandstone types in Kazakhstan and Niger with multiples of 100,000 t U, and the historic Erzgebirge vein type district with over 200,000 t U.

The unconformity-associated and breccia-hosted Mesoproterozoic deposits produce 45% of the world's primary U. In-situ leach (ISL) extraction of sandstone and surficial deposits (e.g. calcrete, deeply weathered granites and groundwater) accounts for some 20% of world U production (World Uranium Mining, 2004). Primary worldwide U production in 1997 was 35,692 t U and this rose to 37,786 t U by 2003. Exploration for new unconformity-type deposits is active globally, particularly at the base of Mesoproterozoic strata such as the McArthur Basin of northern Australia (see comparisons by Gandhi, 1995 and Ruzicka, 1996a), the Aldan and Anabar shields of Russia (Molchanov, 2002) and analogous basins in India (Gandhi, 1995).

Between 1971 and 1990 the contribution of nuclear energy to world electrical power rose from 2 to 17% (Mining

Table 3. Comparison between Athabasca and Thelon basin (after Miller and LeCheminant, 1985; Kyser et al., 2000).

Attribute	Athabasca	Thelon
Graphitic metasediments beneath ore	Distinct	Minor?
Paleo-weathering profile below basal	Deep	Deep
Sub-basins developed via reactivated faults	Yes	Yes
Maximum age of sedimentation (ma)	~1720-1750	1720
Phosphates at base	Minor	Distinct
Aeolian sandstones	Possible	Yes
Arkosic sandstones regionally clay altered	Yes*	Yes
Quartz overgrowths preserve hematite rims	Yes	Yes
Early detrital kaolin in matrix	Yes	No?
Peak diagenetic clays	Dickite + illite	Illite
Peak diagenetic / hydrothermal temperatures	~240º	~200º
Illite incorporates Mg and Fe	Little	Variable
Corroded zircons near ore zones	Local	No?
Regional fresh zircon	Yes	Yes
Extensive Crandallite	Yes	Not reported
K-feldspar and chlorite at 1 Ga	No	Yes
Late vein carbonates from meteoric water	Yes	Yes
Bleaching and clay alteration halos	Yes	Yes
Sandstone-hosted U	Yes	Yes
Basement hosted U	Yes	Yes
Significant deposits	Yes (many)	Yes (few)

* Kyser et al. (2000) consider that evidence for primary depositional feldspar in the Athabasca Basin is very poor; Ramaekers (1990), Ramaekers et al. (2005), Collier (2005) and Bernier (2004) suggest otherwise, particularly for the Fair Point Formation.

Journal, Sept. 18, 1998). In the leading industrial countries, notably the European Union, United States and Japan, nuclear energy now accounts for 20-30% of total electrical requirements and up to 78% in France. Of the world's 438 commercial nuclear reactors, 31 were completed in the past few years and 32 more are under construction. The global share of nuclear-generated electricity could rise to 25% by 2030 (National Post, August 3, 2004).

Spot prices for U_3O_8 in December 2004 reached \$20.70 (U.S.) a pound, 296% higher than they were in 2001, and the highest level in nearly 30 years. Barron's dispatch to the National Post, August 3, 2004 explained that excess product is finally being worked off thereby elevating spot prices. Excess product includes weapons grade material that is being diluted for energy production (Military Warheads as a Source of Nuclear Fuel; Nuclear Issues Briefing Paper 4; July 2003; published by Uranium Information Centre, Melbourne, Australia at <www.uic.com.au>), and domestic stockpiles of U by the various utilities (World Uranium Mining, 2004).

Canadian Unconformity-Associated Uranium Resources

As noted above, current economic Canadian U resources are contained entirely in unconformity-associated U deposits. These are 0.5 to 2 orders of magnitude higher in grade than all other deposits in the world (Table 1, Gandhi, 2005; Ruzicka, 1996a). Past producers of other U deposit types in Canada are all 1-3 orders of magnitude lower in grade, including Paleoplacer conglomerate (Roscoe, 1996) and vein (Ruzicka, 1996b; Ruzicka and Thorpe, 1996). Sandstone U deposits in Canada are relatively small and sub-economic (Bell, 1996). The size and grade distribution of Canadian unconformity U deposits is illustrated in Figure 2 along with comparable deposits in Australia.

In 1997, Canadian production represented approximately 34% of the world total. At that time Canadian U sales were 11,274 t U (29.3 M lbs U3O8) reportedly valued at \$402.25 million (US). Canada's production gradually declined to 28% of the world's primary U. Canadian production may nevertheless reach 50% of world requirements by 2006, based on the projected start-up of production from Cigar Lake (7,000t U) that will be milled at McClean Lake and Rabbit Lake, with continuation of production from McArthur River (7,000t U) milled at Key Lake (World Uranium Mining, 2004; and Canada's Uranium Production & Nuclear Power, Nuclear Issues Briefing Paper # 3, October 2004). Midwest (2,200t U) is another potential new producer in the Athabasca Basin. Given the rise in U prices, other production around the world will probably also increase, in which case Canada's share may only reach ~40% (R. Vance, pers. comm., November 2004).

Nuclear energy fuel endowments in Canada are at the same order of magnitude as fossil fuels (Saskatchewan Mining Association web site: www.ccpg.ca), in absolute terms, as follows. The energy potential of Saskatchewan's U reserves is equivalent to 3.7 billion tonnes of coal or 17 billion barrels of oil. More energy is contained in Saskatchewan's known U reserves than in all known Canadian conventional oil reserves (does not include the Athabasca tar sands). At the current rate of extraction, Saskatchewan's known U deposits will last for more than 25 years. This value only includes known deposits. Exciting new prospects are continually being discovered through intensive exploration.

Geological Attributes

Continental Scale (Geotectonic Environment)

The continental-scale geotectonic environment of significant unconformity-associated U deposits is, in simple terms, at the base of flat lying, thin (less than 5 km) fluviatile strata resting on peneplaned tectonometamorphic complexes in the interiors of large cratons. Cross sections of the Athabasca Basin illustrate these relationships (Figs. 4, 5). The influence of plate or plume tectonics on the origins, diagenesis and mineralization processes of interior Proterozoic basins are considered enigmatic (Ross, 2000). Peneplaned and deeply paleo-weathered basement, and continental sedimentation, implies prolonged stable cratonic environments that persisted before and during the mineralization. This long-held view is modified in light of recent tectonostratigraphic analysis by Ramaekers and Catuneanu (2004) and Ramaekers et al. (2005a), and seismic insights into the deep basement (Hajnal et al., 2005). These provide reasonable and testable modern plate-tectonic hypotheses for the origin of the Athabasca Basin, as follows.

Ramaekers et al. (2005a) deduce that the Athabasca and Thelon repositories of fluvial strata developed accommodation space and subsequent hydrothermal processes by a combination of escape tectonics driven by far field stresses, high heat flow resulting from intrusion of deep mafic magmas (possibly the regional "bright reflector" of Hajnal et al., 1997, 2005; Macdougall and Heaman, 2002) and/or subsidence possibly caused by mantle phase changes associated with relict descending shallow-subduction slabs. The escape tectonic framework (i.e. crustal wedges being squeezed out laterally by converging cratons) is based on evidence of very subtle transtensional to transpressive Mesoproterozoic tectonism, marked by brittle reactivation of a network of regional to secondary and tertiary Paleoproterozoic fault zones in the 1.9 Ga Taltson and 1.8 Ga Trans-Hudson orogens that accommodated ductile transpression during convergence of the Slave, Rae and Superior provinces (e.g. Hoffman, 1988).

Episodic brittle reactivation of one of these fault zones concomitant with emplacement of uraninite ore was documented in Sue Pit by detailed mapping and recording of kinematic indicators (Tourigny et al., 2005). In the McArthur River area, structural and stratigraphic analysis of drill core (e.g. McGill et al. 1993) documented 40-80 m of reverse brittle offset along the P2 Fault. Such brittle offsets affected not only the basal unconformity, but also transect early hydrothermal alteration features such as silicification and control the location of ore. Regional (Hajnal et al. 2005) to detailed (Gyorfi et al., 2005) analysis of seismic data has shown the deep listric nature of the P2 Fault and its geometric relationship to previous ductile faults that were developed during fold and thrust tectonics of the Hudsonian orogeny. Detailed sedimentologic and stratigraphic analysis (Bernier, 2004, Long, 2005, Yeo et al., 2005b) has shown that such

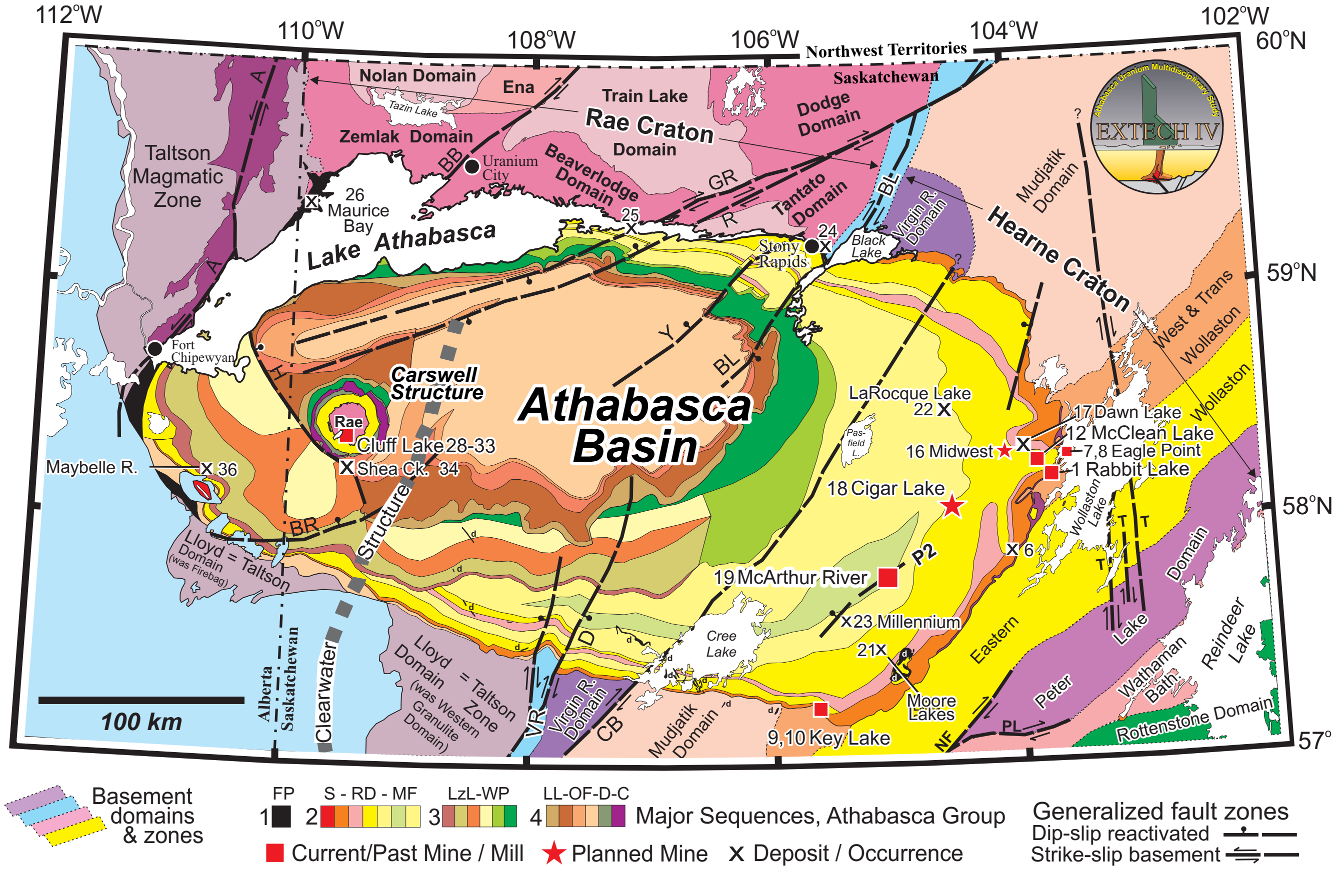

FIG. 4. Regional geology of the Athabasca Basin region in northern Saskatchewan and Alberta. Deposits and significant occurrences are numbered as listed in Table 1 and detailed in Gandhi (2005). After Card et al. (2003, 2005a, b), Portella and Annesley (2000a), Ramaekers et al. (2005b) and Thomas et al. (2002). Sequences (from Ramaekers et al., 2005b) are summarized in Table 4. Generalized fault zones (these include multiple ductile movements before deposition of Athabasca Group and brittle transcurrent and dip-slip movements during and after deposition) are: A = Allan; BB = Black Bay; BL = Black Lake; BR = Beatty River, CB = Cable Bay, D = Dufferin; GR = Grease River; H = Harrison; NF = Needle Falls; PL = Parker Lake; P2 = "P2 Fault" at McArthur River; R = Robillard; T = Tabbernor; VR = Virgin River Shear Zone (Dufferin is one named fault of many in VR); Y = Yaworski. Cross Sections of Figures 5A and 5B are located along dotted lines labelled NW - SE (along the basin axis) and E - W (south of Key Lake). The western Wollaston domain and the Wollaston - Mudjatik transition of Portella and Annesley (2000a, b) are combined here as "West + trans. Wollaston".

faults are spatially related to pre-Athabasca Group paleotopographic features such as paleo-valleys and minor fault scarps. Stratigraphic and seismic analyses have also documented the growth of basement highs and tilting of the basin during sedimentation. The highs, commonly termed "quartzite ridges" are interpreted (Gyorfi et al., 2002, 2005) as compressional pop-up structures, with brittle basement blocks of various sizes having been squeezed upward into the unconsolidated overlying strata. The tilting is interpreted (Ramaekers et al. 2005a) as a local manifestation of increased accommodation space on one side while uplift and erosion took place on the other side.

On the regional scale, a series of stratigraphically and temporally constrained isopach maps, facies changes and linked paleocurrent data also show spatial relationships with these bounding faults (Ramaekers et al., 2005a, b), including the development of various sub-basins through time, and the hinge lines that separated these sub-basins. Parts of the basin that were at one time paleotopographic highs later became lows that accumulated thicker strata. The polarity of proximality and distality was at times east to west, and at other times south to north. From these indications of tectonic activity, it is apparent that second- and third-order derivatives of subtle continent-scale compressive tectonics influenced the movements of basinal and basement fluids, some of which precipitated world class ore deposits and variously altered both basement and sedimentary cover.

Geological Attributes on the Scale of Metallogenetic Districts of Canada

The Athabasca Basin (Fig. 4) is by far the most significant U metallogenic district in Canada, in terms of known deposits and production (Table 1). Existing deposits and prospects (Table 1) are undergoing intense new exploration, re-evaluation and development to establish economic reserves and/or increase potential production. The Thelon Basin is similarly experiencing renewed exploration interest, and contains the only other significant prospects of the unconformity-associated type (Fig. 3, Tables 1 & 2). Various other Mesoproterozoic to slightly older red-bed basins in Canada (e.g. Sibley and Otish, Figure 1) are being re-evaluated for their potential to host deposits of this type.

In the Athabasca Basin (Figure 4) the great majority of mines and prospects are located where the Athabasca Group unconformably overlies the western Wollaston and Wollaston-Mudjatik transition basement domains, however,

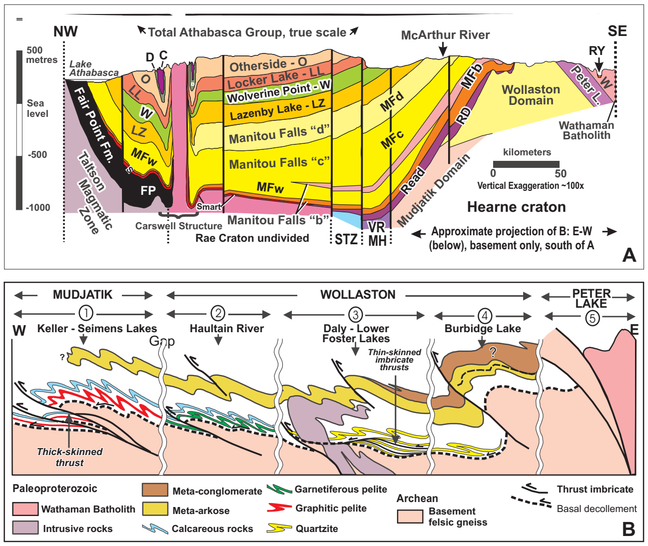

FIG. 5. A. Lithostratigraphic cross section of the Athabasca Basin, after Ramaekers (1990) and Ramaekers et al. (2005b). True 1:1 scale is shown at top. Stratigraphic units in the vertically exaggerated section are formations except those starting with MF are members of Manitou Falls Formation; MFw is Warnes Member. Basement domains are diagrammatic. Line of section is shown as NW-SE in Figure 4. B. Structural cross section adapted from Gyorfi et al. (2005) after Tran (2001) illustrates structural geometry of the Wollaston-Mudjatik transition zone, south of Key Lake that is representative of the rocks that underlie the most productive area of the eastern Athabasca Basin. Line of section is shown as W-E in Figure 4.

significant mined deposits and prospects of the Cluff Lake camp in the Carswell Structure and new prospects intersected by drilling at Maybelle River (Kupsch and Catuneanu, 2005) and Shea Creek (Rippert et al., 2000) demonstrate the potential for unconformity associated U deposits in the western part of the Athabasca Basin.

Metallogenic Districts of Canada

Active mining of unconformity-associated U deposits is restricted to the Athabasca Basin in Canada. Other metallogenic districts include the Thelon Basin of NWT-Nunavut and the Hornby Bay basin on the northeast corner of Great Bear Lake. Past-producing vein U districts of the Beaverlodge Camp (Uranium City area north of Lake Athabasca) (Ruzicka, 1996b) may be exhumed unconformity-associated deposits (Mazimhaka and Hendry, 1989). The Great Bear Lake (Port Radium) area (Ruzicka and Thorpe, 1996) may also be related to the sub-Hornby Bay Group unconformity, but neither this nor the Beaverlodge camp are treated in detail here or included in the database. The Beaverlodge deposits also represent a source of re-cycled U for the Athabasca Basin deposits. The Otish Mountains area of Quebec is another site of exploration for unconformity-associated U (Ruzicka, 1996a).

<u>Time and Space Distribution of Unconformity-associated Uranium Districts</u>

Unconformity-associated U deposits in the Athabasca and Thelon basins are constrained in age by recent detrital and diagenetic geochronology of stratigraphic sequences above the unconformity, linked with existing geochronology of the deposits. Rainbird et al. (2005) estimate that sedimentation began in the Athabasca Basin at about 1740-1730 Ma, considering metamorphic ages on titanite as young as 1750 Ma in basement rocks (Orrell, et al., 1999). The Barrensland Group of Thelon Basin also has a maximum age of 1750-1720 Ma (Miller, et al., 1989; Rainbird, et al., 2003a) based on ages of early diagenetic phosphatic material in basal strata.

The upper ages of these two groups are weakly constrained. Rainbird et al. (2005) have dated internal tuffaceous units in the third sequence of the Athabasca Group (Wolverine Point Formation) at 1644±13Ma (U-Pb), close to previous approximate U-Pb dates of >1650-1700 Ma on diagenetic fluorapatite in Fair Point and Wolverine Point formations by Cumming et al. (1987). Sequence 4 is capped by organic-rich shale of the Douglas Formation that appears to be about 100 Ma younger (Creaser and Stasiuk, 2005) and carbonate (Carswell Formation) whose upper age is unconstrained. Uranium deposits could have formed before either of these times, and the fluorapatite ages of >1650-1700 Ma suggest a basin-wide diagenetic/hydrothermal event at about that time. Available geochronology of Athabasca U deposits records one or two main hydrothermal ore-related events within the basin at circa 1500 and 1350 Ma that were overprinted by further alteration and U remobilization events at approximately 1176 Ma, 900 Ma and 300 Ma (Hoeve and Quirt 1984; Cumming and Krstic 1992; Fayek et al., 2002a). This implies that the U deposits began to form while sediment was still accumulating in the Athabasca Basin, after early diagenesis and during late, high-temperature diagenesis with a remarkable time span of at least 100 Ma, and possibly more than 200 Ma.

Unconformity U deposits beneath the Thelon basin likely formed at about the same times as those of the Athabasca Basin (Kyser et al. 2000), although the oldest date obtained thus far is 1400 Ma on the Kiggavik deposit (Fuchs and Hilger 1989). Hornby Bay and Elu basins are stratigraphically comparable with the Thelon and Athabasca basins, and have been explored for unconformity-related deposits. Hornby Bay and Elu basins differ from the Athabasca and Thelon basins in their continent-margin rather than intracratonic settings, and in lacking known unconformity-associated deposits. Hornby Bay Group does, however, host a disseminated sandstone-type deposit, which is the only significant example of this deposit type in the Proterozoic of Canada (Bell, 1996).

Near the Hornby Bay basin, shear zone-hosted U-Ag-Co-Cu-As veins of the Great Bear Lake area were the original source of Canada's radium and U production, and later produced significant Ag (Ruzicka and Thorpe, 1996). Ruzicka and Thorpe (1996) interpreted these veins as having been formed by hydrothermal processes possibly as late as 1500 to 1400 Ma (U-Pb on pitchblende, Jory 1964), "soon after the Hornby Bay Group was deposited". The upper Narakay Volcanics overlying the Hornby Bay Group have been dated by the U-Pb method at 1663 $^{+}$/- 8 Ma by Bowring and Ross (1985). Ruzicka and Thorpe (1996) cited older 1775 to 1665 Ma ages by the Pb-Pb method on the same veins however such ages are based on many assumptions and are therefore suspect. If the 1500 to 1400 Ma U-Pb ages of these deposits are correct, then these deposits are exhumed candidates for the unconformity-associated U deposit type - they have the same structural characteristics as basement-hosted unconformity U deposits but slightly different associated elements, in particular high silver.

The Beaverlodge Domain north of Lake Athabasca produced significant U (Table 2) from shear-zone-hosted pitchblende veins in the basement with uraninite ages of about 1780 Ma (Koeppel, 1967), and yet these have also long been considered early analogues or exhumed roots of unconformity associated deposits beneath red beds of the fault-controlled Martin Basin (1630-1830 Ma; Tremblay, 1972; Maximhaka and Hendry, 1989; Card et al., 2005b). The Gunnar deposit of the Beaverlodge camp is an exception (see episyenite, below).

Other regions in Canada have partially favourable geological settings for unconformity associated U deposits, e.g. the Paleoproterozoic Otish Mountains Group (with some U prospects, Ruzicka, 1996a) and younger Paleoproterozoic red bed sandstone portions of the Hurwitz Group (the upper part of which is now dated as <1.9 Ga by U-Pb on detrital zircon by Davis et al., 2005) and upper Huronian Cobalt Group. These basins are red-bed siliciclastic sequences, but may be too old to have experienced the same atmospheric and tectonic environment as the Athabasca Basin. These have been variably tectonized, particularly the upper Hurwitz that is dismembered and essentially equivalent to the Wollaston basement supracrustal belt beneath the Athabasca Basin. The Sibley Group is relatively flat lying but a very large proportion of it is intruded by diabase, and it may be a bit too young.

<u>Expressions of Favourable Geological Environments or Events at the District Scale</u>

First-order favourable environments are the unconformities between relatively flat-lying and intracontinental, unmetamorphosed, late Paleoproterozoic to Mesoproterozoic, siliciclastic fluvial red-bed strata and underlying Paleoproterozoic metamorphosed supracrustal rocks. Especially favourable are the intersections of these unconformity surfaces with underlying peneplaned fold and thrust belts (e.g. Lewry and Sibbald, 1979) that include graphitic metapelites containing naturally high U contents, including small U-prospects, such as the Wollaston Group (Yeo and Delaney, 2005) and the Amer Group (Miller and LeCheminant, 1985). Such belts typically include regional zones of basement ductile faults and fracture zones. These faults are susceptible to brittle reactivation and are important for basin development (Kerans et al., 1981; Ramaekers, 1990; Ramaekers et al. 2005a) as well as later mineralizing processes (Hoeve and Sibbald 1978; Ramaekers et al. 2005a; Tourigny et al. 2005).

Many workers, e.g. Annesley et al. (1997), Cuney et al. (2003), Freiberger and Cuney (2003) and Hecht and Cuney (2003) have emphasized the favourability of basement domains that have U-enriched rock types such as pegmatites, metapelites and radiogenic granitoid rocks. Such domains, also rich in potassium (K), thorium (Th) and rare earth elements (REE) may have selectively contributed U to the prospective basins through a variety of mechanisms as described below under Genetic Models.

Prospective Proterozoic basins in Canada and Australia are typically underlain by extensive paleo-regoliths that have also been altered since deposition of the overlying strata (e.g. Cecile, 1973; Fraser et al., 1970; Gall, 1994; Hoeve and Quirt, 1984; Kyser et al., 2000; McDonald, 1980, 1985; Miller et al., 1989; Ramaekers, 1990). The paleo-weathered basement immediately below the Athabasca Group has a vertical profile ranging from a few centimetres up to 70 metres thick, with much deeper pockets and slivers developed along fault zones. The following description is after Macdonald (1980, 1985). The regolith is variably affected by diagenetic iron reduction resulting in a bleached zone at the top, interpreted as hematite removal from the upper "red zone" of the paleo-weathered basement section. This bleached zone, always present immediately below the unconformity, is composed of buff-coloured clay and quartz. It crosscuts and therefore post-dates the red zone beneath. Below this, the upper portion of the regolith profile exhibits a strong red hematitic alteration that grades into underlying greenish chloritic alteration. In profiles developed on the basement meta-arkose, a white zone of white clay replacement of feldspars and mafic minerals separates the red and green zones. A downward progression from kaolinite to illite and chlorite is common through the regolith profile.

Discussion continues regarding the respective contributions of paleo-weathering, diagenesis and hydrothermal processes to the alteration preserved at the unconformity. Cuney et al. (2003) stated that the regolith may represent a redox front controlled by the degree of percolation of the oxidized diagenetic brines in the basement, and that the lack of Ce-anomalies is not in favour of a lateritic origin. It is here preferred that the red-green alteration was indeed regional, a result of lateritic weathering as proposed by Macdonald (1980), and was overprinted by hydrothermal alteration (the white zone of Macdonald) that increases in intensity close to and is genetically related to the bleaching alteration associated with U deposits (Macdonald, 1985). This interpretation is based on field relationships noted by Macdonald (the white zone transects down across the red and green along fractures).

Macdonald's observations are supported by evidence of red hematitic environment during deposition of the basal members of the Athabasca Group, in particular the Read Formation (formerly MFa) in the eastern part of the basin (Ramaekers et al., 2005b). In particular, Yeo et al. (2005a) summarized various observations of oncoidal structures preserved in red mudstone and microbial laminae preserved in silicified conglomerate. These structures are made up of strongly hematitic but delicate fine laminae that must be primary, demonstrating an iron-rich and highly oxidized early environment of sedimentation for the Read and Smart Formations. Similar red mudstones are present at the base of the Fair Point Formation, but are thinner and lack oncoids.

Nevertheless the above evidence is not conclusive of paleo-weathering according to some authors (e.g. Cuney et al., 2003) who propose that the reddening took place during diagenesis. One example of conclusive evidence of paleosol development would be incorporation of blocks of paleosol with red-green zonation preserved in the basal conglomerate. However the white clay alteration along the unconformity zone extends up into most of the basal siliciclastic units and obliterates any such zonation that might have been preserved in lithic fragments. More research is required on this issue. In any case, the red-green basal unconformity transition represents a regional redox boundary that was variably overprinted by regional to local diagenetic and hydrothermal processes. Strong field-controlled geochemical research involving complete sections from the Athabasca and Thelon groups down into fresh basement are required at locations distal and proximal to ore, in order to test these various hypotheses.

The siliciclastic strata that overlie the paleo-regolith and host the U deposits are thoroughly oxidized terrestrial red bed sequences with very long and complex diagenetic histories (Table 3). Each pulse of sedimentary accommodation development in the host basins was not only associated with the accumulation of new strata, but also with changes in basin configuration, re-activation of the various growth faults at hinge lines, and concomitant changes in, or renewals of, hydrogeological and hydrothermal systems (Ramaekers et al. 2005a). These systems altered sedimentary and metamorphic minerals to clays, dissolved significant amounts of the rock locally to regionally, locally silicified strata up to 100% and/or introduced K as illite [$(H_3O,K)_y(Al_4.Fe_4.Mg_4.Mg_6)(Si_{8-y}Al_y)O_{20}(OH)_4$] and boron (as dravite). The illite alteration zones transect stratigraphy and therefore the introduction of K is not a result of altered primary arkosic beds). Such alteration not only transected large volumes of strata, it also modified the regional basement-cover redox boundary. The degree of association of these changes with U mineralization is addressed under deposit-scale alteration below. Many of these diagenetic changes are basin-wide: depositional clays were kaolinite [$Al_2O_3.2SiO_2.2H_2O$]; now the regional clay mineral is a mixture of dickite [$Al_2Si_2O_5(OH)_4$], a polymorph of kaolinite, either "still there" altered from original kaolinite or later formed) and illite in the Athabasca Basin (Wasyliuk, 2002). The regional clay is illite in the Thelon Basin (Kyser et al., 2000).

In the Athabasca Basin, in addition to the above-described regolith and hydrothermal alteration of basement below the unconformity, two other types of regional-scale alteration have been distinguished:

1) Basin-wide pre-ore diagenetic sandstone alteration, and

2) Sub-basin-scale ore-related alteration halos.

One of the earliest recognizable diagenetic events in the Athabasca basin is a pre-ore quartz overgrowth (Q1 event) that encapsulates hematite-coated detrital quartz grains. In the eastern Athabasca Basin the original clay matrix of the sandstone, presumed to be dominantly kaolinitic with some(?) detrital feldspar (Hoeve and Quirt 1984; Ramaekers 1990; Quirt 2001), has been diagenetically altered mostly (but not completely) to dickite (Earle et al., 1999, Quirt 2001; Wasyliuk 2002), a higher crystallinity polymorph of kaolinite. Kyser et al. (2000) hold that evidence for primary detrital feldspar is lacking in the Athabasca Basin, and that all feldspar was altered to clay before incorporation in the basin fill. Macdonald (1980) had previously favoured this interpretation as well. They did however note the local preservation of microcline in early-silicified basal units of the Thelon Formation. Our view is that, given the 600 million year span of high temperature diagenesis in the Athabasca Basin, the near-complete lack of preserved feldspar (except in Fair Point Formation) does not preclude its primary deposition in the Manitou Falls and younger formations.

A variation in the regional background "dickitic" pattern was noted by Earle and Sopuck (1989) in the south-eastern part of the basin where a large illite anomaly forms a corridor, 10-20 km wide, that extends for 100 km northeast from Key Lake (Figure 6). Earle et al. (1999) describe the illitic alteration at Key Lake in more detail. The axis of this regional illite anomaly also contains sub-parallel linear zones of anomalous chlorite and dravite. This anomaly also encompasses all known U deposits and prospects in the southeastern part of the basin, notably Key Lake, P-Patch (4 km East of Key Lake), McArthur River, BJ (just SE of McArthur River) and the Millennium prospect.

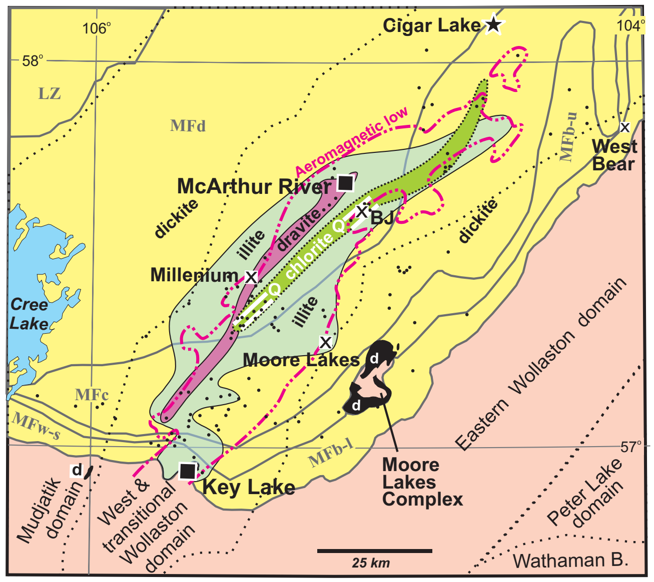

FIG. 6. Lithogeochemical map of the south-eastern part of the Athabasca Basin, showing regional illite, chlorite and boron anomalies in the surficial material and outcrops of the Athabasca Group. Also projected to surface are areas of low magnetic susceptibility in basement rocks, and basement quartzite ridges (white bars labelled Q). After Earle and Sopuck (1989). Basement units (no pattern) are dotted beneath the Athabasca Group (grey). Stratigraphic units in the Athabasca Group are: MFb-l, lower Bird Member of Manitou Falls Formation; MFb-u, upper Bird Member; MFc, Collins Member; MFd, Dunlop Member; LZ, Lazenby Lake Formation. "d" = diabase. Squares are mines and mills, star is planned mine, Xs are prospects.

Basement rock compositions may have influenced the lithogeochemistry of the overlying sandstone because several spatial associations have been noted. These, however require quantitative testing. One apparent spatial association is the above-described regional illite (+chlorite + dravite) anomaly that overlies a broad aeromagnetic low region, 5-20 km wide, wherein the underlying Wollaston supracrustal gneisses include abundant metaquartzite and metapelite units. The illite anomaly is expressed as relative K anomalies in ternary K-U-Th airborne and ground spectral gamma ray surveys (Shives et al. 2000). Another spatial association is related to the distribution of clay minerals in various sequences of the Athabasca Group. Although these siliciclastic strata are now mineralogically dominated by quartz, petrographic evidence suggests that significant labile minerals were originally incorporated in the detritus and were pervasively converted to clay minerals (Wasyliuk, 2002). An alternative hypothesis is that the paleo-weathering regime was so intense that only clays and quartz were transported into the basin (Macdonald, 1980). This hypothesis certainly is supported by the nature of the Read Formation but not necessarily the overlying Manitou Falls Formation or older Fair Point Formation (Ramaekers et al., 2005b).

The regional distribution of dickite with minor illite has been proposed in part to reflect a combination of unique sandstone stratigraphy as well as "deep diagenesis" involving alteration of stratigraphically controlled detrital clay by formational brines at temperatures exceeding 100°C (Quirt, 2001). Nevertheless, translating such alteration results into experiments with clay mineral "stratigraphy" produced results in the area of the structurally complex Carswell impact feature (e.g. Hoeve et al., 1985) that are completely inconsistent with those based on primary grain-size parameters (e.g. Collier 2002, 2003, 2005; Ramaekers et al. 2005b, Yeo et al., 2001a, b). In order to maintain a consistent basin-wide stratigraphic framework it is necessary to consider alteration mineralogical data only as a post-depositional effect that commonly sharply transgresses primary stratigraphic units defined on framework-textural attributes. This is vitally important for exploration, because it is this very transgressive nature of clay mineralogy that records fluid movements and hence vectors to the paths of ore-forming fluids.

Evidence of detrital labile minerals is most apparent in the Fair Point Formation which is restricted to the western part of the basin, is characterized by abundant matrix clay and has a likely detrital volcanic component as suggested by hematitic cobbles with relict volcanic texture (Ramaekers et al. 2005b). Other evidence of labile minerals is in the Bird Member of Manitou Falls Formation, in which mafic heavy mineral laminae are interpreted from thin hematite-rich layers that are located at the base of conglomeratic beds, wrap around pebbles to emphasize sieve texture, and delicately outline trough cross laminae (Mwenifumbo and Bernius, 2005). These laminae now retain only zircon and quartz framework grains surrounded by a complex matrix of hematite, Th-rich aluminum phosphates and clay minerals, some of which occupy framework gaps and are the shape and size of adjacent quartz grains (Mwenifumbo and Bernius 2005; Mwenifumbo et al. 2005).

A third spatial association is derived from integrated borehole geophysical and stratigraphic data: complex aluminum phosphate minerals correlate with Th anomalies and grain-size parameters, and appear to be systematically depleted in U with respect to other elements (Mwenifumbo et al., 2005; Mwenifumbo and Bernius, 2005). These correlations are observed in the western part of the Athabasca Basin at Shea Creek (Mwenifumbo et al., 2000) but are most strongly developed in conglomerate beds of the Manitou Falls Formation that are restricted to the eastern part of the Basin. The Th anomalies in these conglomerates are evident regionally in drill core (Mwenifumbo and Bernius 2005). Th is also anomalous in regional airborne spectral -ray data that map Th anomalies in ternary K-U-Th plots (NATGAM database, Radiation Geophysics Section). These Th anomalies coincide with the distribution of the Bird Member of Manitou Falls Formation (Campbell et al. 2002). Aluminum phosphates are present throughout the clastic units of the Athabasca Group (M. Cuney, pers. comm. 2005), but these are most evident in the Manitou Falls Formation (Mwenifumbo and Bernius 2005). Sharp peaks in phosphate abundance are spatially associated with relict black sand laminae (now amorphous hematite cement) that strongly resemble paleo-placer heavy mineral accumulations and imply that some of the heavy minerals (especially monazite but not zircon) were preferentially altered to aluminum phosphates while releasing U in solution (Mwenifumbo and Bernius, 2005).

Geological Attributes on the Scale of Deposits

Local Geological Settings and Controls on the Siting of Ores

A close relationship between local faults and U accumulations has been known since exploration and development of the Rabbit Lake deposit (e.g., Hoeve and Sibbald, 1978; Hoeve et al., 1980) and have been an increasingly important part of the exploration framework ever since (e.g. Baudemont and Rafini, 2000). Tourigny et al. (2005), through detailed mapping of the Sue Pit at McClean Lake, demonstrated a direct spatial and temporal association between transpressional reactivation of ductile Hudsonian basement fault zones and deposition of uraninite in dilatant jogs, whose geometries predict the overall deposit geometry. Uraninite pods and lenses at the north end of Sue Pit are located in the basement very close to the unconformity, but rake toward the south, such that the ore is located well below the unconformity at the south end of Sue C Pit. Harvey and Bethune (2005) noted similar structural associations in the Deilmann Pit at Key Lake.

Irregularities in the basal unconformity are an important corollary to the demonstrated reactivated fault systems and have also been known for some time. One such irregularity is spatially associated lows in the unconformity surface, as documented by Harvey and Bethune (2005) through isopach analysis at Key Lake. Such basement lows were shown to be in part true paleo-valleys, through sedimentological analysis by Long (2005) at Sue Pit, Collier and Yeo (2000) at Dielmann Pit, Bernier (2004) at McArthur River, and Yeo et al. (2005b) at McArthur and Wheeler rivers. Another irregularity is basement ridges, long thought to have been actual paleotopographic ridges that existed before sedimentation (e.g. Earle and Sopuck, 1989). Analysis by Bernier (2004) and Yeo et al. (2005b) further demonstrated that growth faults, paleo-valleys and basement highs developed actively before, during and after sedimentation. Parallel coordinated high-resolution seismic imaging by Gyorfi et al. (2005) at McArthur River has come to essentially the same conclusions, and demonstrated that such features can be mapped accurately along multiple transects. Some of these are related to the specific faults that host the uraninite deposits and are the focus of zoned alteration halos (see below), various aspects of which can be mapped by mineralogical, seismic, magnetotelluric and gravity analysis. Some basement highs are tectonic escape structures that developed during and after sedimentation while the region was undergoing compression, at the same time as accommodation space was being developed and filled by fluvial siliciclastic detritus (Bernier 2004, Gyorfi et al., 2005, Yeo et al. 2005b).

Graphitic basement units associated with the fault zones are an essential component of deposit architectures in the Athabasca and Thelon basins. Underlying the eastern Athabasca Basin, graphitic units are stratigraphically low in the metasedimentary zones 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, 2005). Similar graphitic units are found beneath western Athabasca deposits in the Maybelle River (Pana et al., 2005) Carswell Structure, and Shea Creek areas where they are considered as supracrustal components of the Taltson Magmatic Zone (Card et al., 2005a; Brouand et al., 2003; Rippert et al., 2000).

Graphitic units underlying deposits in the Thelon Basin are identified as part of the Amer Group (Miller and LeCheminant, 1985). As shown by Tourigny et al. (2005) for the Sue Pit at McClean Lake (deposit 12 in Fig. 4), graphitic metapelites are zones of weakness along which faults have repeatedly propagated. The graphitic units are also strong conductors and serve as excellent exploration targets for electromagnetic methods. These units have also been regarded as a key genetic component in geochemical process models for unconformity-associated U, along with considerable controversy over whether (McCready et al. 1999), or not (Wilson et al., 2005) they could have supplied sufficient organic reductants to directly and quantitatively precipitate the world-class concentrations of uraninite that characterize the Athabasca Basin. Electrochemical processes are another way in which graphitic units may have focused uraninite precipitation from hydrothermal fluids, by serving as anodes of natural electrical systems.

Deposit Size, Morphology and Architecture

Deposit tonnages and grades are summarized Table 1 and Figure 2. The physical size and shape varies considerably: individual super-high-grade ore pods of massive uraninite (referred to as "Zones") at the world-class McArthur River deposit reach 100 m or more in vertical extent, and 50 m in cross section, with mining grades in the order of 20-25% U. High-grade lenses in Sue C pit are in the order of 1-2 m thick and 3-5 m in vertical dimension, situated en-echelon in a zone hundreds of metres in strike length and extend down-depth for tens of metres. These pods are bounded by sheared and brecciated graphitic schist that contains other smaller lenses of the same material, forming an envelope of low-grade ore that can be blended with high-grade ore during mining and milling if necessary. Typical mining grades of these deposits are in the order of 0.5 to 2% U. The Cigar Lake deposit is completely different in shape. It comprises three upward convex lenses ranging from 50 to 100 m across, reaching maximum thickness of about 20 m and each being situated at the unconformity along a total strike length of about 2 km. Only the eastern two lenses, with a total strike length of about 600 m, are being groomed for Phase 1 production of some 496,780 tonnes at an average grade of about 20.7% U_3O_8 (Andrade, 2002).

These deposits are, in simple terms, sub-horizontal cigar-shaped to elongated skewed "T" shaped, but details of morphology and architecture are highly varied, ranging between two end member styles that reflect both stratigraph-

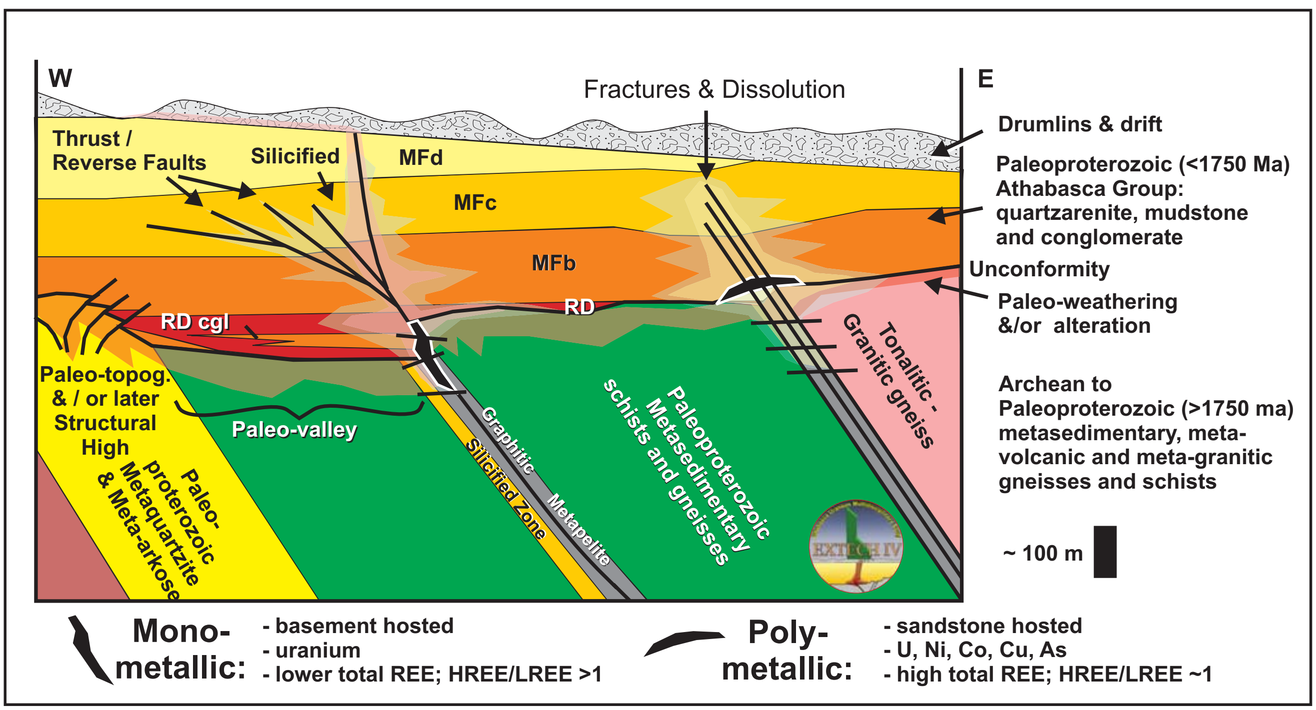

FIG. 7. Generalized geological elements of mono- and poly-metallic unconformity-associated U deposits in the eastern part of the Paleoproterozoic Athabasca Basin. After Hoeve and Quirt (1984), Hoeve and Sibbald (1978), McGill et al. (1993), Ruzicka (1996a), Sibbald et al. (1976), Thomas et al. (2000) and Tourigny et al. (2005). This is an empirical geological framework model illustrating two end-member styles of ore, between which a complete spectrum of styles are known, even within single deposits and deposit groups.

ic and structural control (Hoeve and Quirt 1984; Sibbald 1985; Thomas et al. 2000) (Figs. 4, 5, 7, 8):

1) fracture-controlled, dominantly basement-hosted (e.g. McArthur River, Rabbit Lake, Eagle Point, McClean - Sue C, Dominique-Peter, Raven and Horseshoe), and
2) clay-bounded, massive ore developed along and just above the unconformity in the overlying silici clastic Athabasca Group (e.g. Cigar Lake, Key Lake, Collins Bay A, B and D zones, other McClean deposits, Midwest and Cluff Lake D zone).

The fracture-controlled basement ore typically occupies steeply to moderately dipping shear, fracture and breccia zones, which in places extend 400 m into basement rocks below the unconformity. Disseminated and massive uraninite/pitchblende occupies fractures and breccia matrix, which commonly have grades between 1% and 3% U_3O_8 (0.8 to 2.5 %U). The major exception is McArthur River where production grades are in the order of 20-25% U_3O_8 (16 to 20 %U), and the monomineralic ore extends well up into silicified sandstone, still constrained by the host P2 fault zone and silicified sandstone. In contrast, clay-bounded ore is developed along the basement-sandstone unconformity and forms elongate, pipe-like, and cigar-shaped ore bodies typically characterized by a high-grade core (1-15% U_3O_8) and surrounded by a lower grade halo (<1% U_3O_8). Most of the ore bodies have root-like extensions into the basement. In places, uraninite also extends up into the overlying siliciclastic strata, along cataclastic breccia and fracture zones. Isolated above them are small "perched" occurrences of disseminated uraninite that are rarely of ore grade but are good indicators of potential ore at depth and typically are regarded as "young" remobilised primary ore.

<u>Ore Mineralogy, Chemistry and Zonation</u>

These deposits are essentially massive to disseminated uraninite. The field term "pitchblende" is used to refer to the commonly sooty, cryptocrystalline, botryoidal form of uraninite. The sooty appearance of uraninite is in part due to crushing, milling and alteration associated with multiple post-ore deformation. Microscopically and crystallographically the ore is all uraninite. Much of the ore preserves coarsely crystalline forms of uraninite, and systematic petrographic study produces consistent paragenetic sequences (Wilson et al., 2005). Unconformity associated U deposits are commonly referred to as either monometallic (also known as simple) or polymetallic (complex) on the basis of associated metals (Ruzicka 1989, 1996a; Thomas et al., 2000) (Fig.7).

Polymetallic deposits are typically hosted by sandstone and conglomerate, situated within 25-50 m of the basement-sandstone unconformity. At Cigar Lake the unconformity assemblage hosts the ore. This consists of hydrothermally altered paleo-regolith and basal sandstone-conglomerate. Polymetallic ores are characterized by anomalous concentrations of sulfide 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.

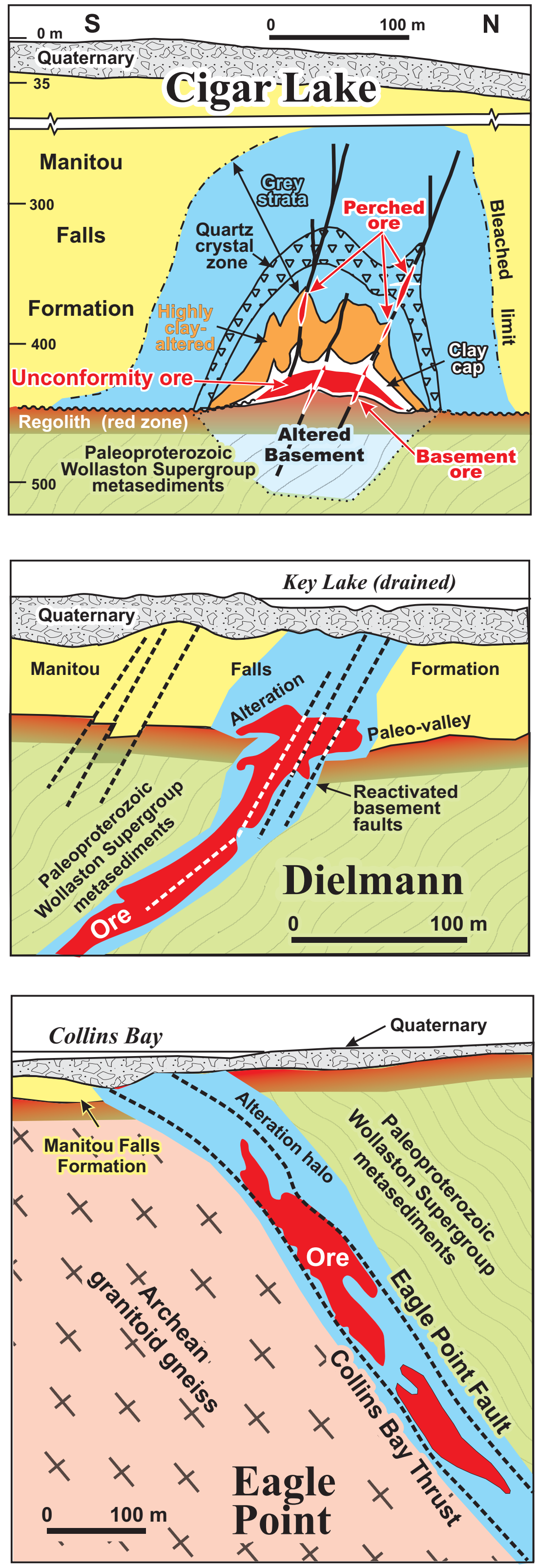

FIG. 8. Examples of three main sub-types of unconformity-associated U deposits, after Thomas et al. (2000) and Andrade (2002). Eagle Point (original deposit mined out but trend still being mined, LeMaitre and Belyk, oral presentation, 2005) is wholly basement hosted; Dielmann (mined out) at Key Lake included both basement-hosted and unconformity ore; Cigar Lake is dominantly unconformity ore with minor basement hosted lenses and perched ore in the overlying Manitou Falls Formation.

In contrast, basement-hosted deposits, typically located greater than 50 m below the unconformity, as well as small perched sandstone deposits or lenses, are referred to as monometallic or "simple", because they contain only traces of metals other than U and minor Cu. The remarkable McArthur River deposit is a special example of a super-high grade "simple" deposit that extends from just above the unconformity (~500 m below surface) to more than 90 metres below the unconformity (Jamieson and Spross, 2000; Jefferson et al., 2002; McGill et al., 1993; Thomas et al., 2000). This "monometallic" deposit does however contain minor galena, pyrite, chalcopyrite, Ni-Co sulpharsenides and gold (Gandhi, 2005).

Alteration Mineralogy and Geochemistry

Alteration mineralogy and geochemistry of Australian and Canadian unconformity associated deposits and their host rocks in the Athabasca and Thelon basins, and the Kombolgie Basin have been compared by 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 and Quirt (1984), and Wasilyuk (2002) has set the modern framework of exploration clay mineralogy. Intense clay alteration zones surrounding deposits such as Cigar Lake 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). The similarities and differences of geological, diagenetic and hydrothermal alteration histories in the Athabasca and Thelon basins are summarized in Table 3.

Alteration halos are developed mainly in the siliciclastic strata overlying "egress type" unconformity-associated U deposits, and range between two distinctive end member types as illustrated in Figure 9: 1) quartz dissolution + illite and 2) silicified (Q2) + kaolinite + dravite. Deposits in the northern part of the eastern Athabasca Basin characteristically underwent quartz corrosion (Hoeve and Quirt, 1984) with volume losses locally exceeding 90% (Percival, 1989; Andrade 2002; Cuney et al., 2003), whereas mineralization

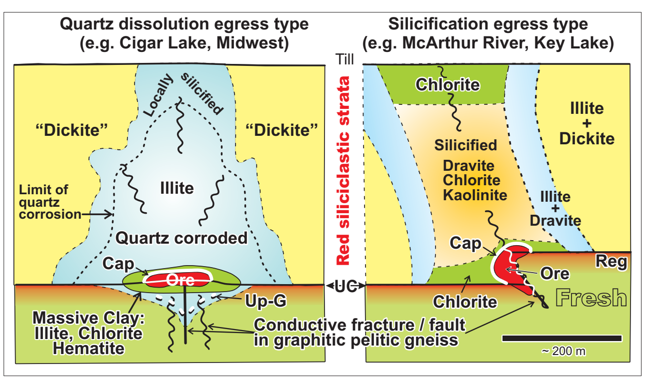

FIG. 9. Two end-member sandstone alteration models for egress-type deposits in the eastern Athabasca Basin (modified from Matthews et al. 1997, Wasyliuk, 2002 and Quirt, 2003). UC = unconformity, Reg = regolith profile grading from red hematitic saprolite down through green chloritic alteration to fresh basement gneiss, Up-G = upper limit of graphite preserved in footwall alteration zone. Cap = secondary black-red earthy hematite capping ore bodies. Fresh = fresh basement: Fe-Mg chlorite biotite paragneiss.

in the McArthur River area is dominantly represented by the silicification end member with very localized intense quartz corrosion and little apparent volume loss (Matthews et al. 1997). Around the Dielmann orebody at Key Lake, silicification is minor, but kaolinite and dravite are superimposed on a regional illite zone that extends south from McArthur River (Harvey and Bethune, 2000).

Ore-related illitic alteration is reflected in anomalously high illite proportions (Hoeve et al. 1981a, b, Hoeve and Quirt 1984, Percival et al., 1993) and resultant anomalous K_2O/Al_2O_3 ratios in the sandstone (Earle and Sopuck 1989). Sudoite, an Al-Mg-rich di-tri, octahedral chlorite (Percival and Kodama, 1989), is present in representatives of both alteration types. Similarly, local silicification fronts are also present in some of the larger de-silicified alteration systems, e.g. Cigar Lake (Andrade 2002). Deposit-related silicification (Q2 event), recorded notably by drusy quartz-filled fractures (McGill et al. 1993) and pre-ore tombstone-style silicification (Q1; Yeo et al. 2001b; Mwenifumbo et al., 2004, 2005), are particularly intense in Athabasca sandstone sequences above or proximal to basement quartzite ridges (previously thought to be pre-Athabasca Group paleotopographic features, now interpreted as mainly syn- and post-sedimentary tectonic uplifts, Bernier, 2004; Gyorfi et al., 2005; Yeo et al., 2005b). The tombstone silicification of the Athabasca Group preserves diagenetic hematite and dickite (Mwenifumbo et al. 2005) and microbial laminae (Yeo et al. 2005a) very close to the McArthur River deposit. 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 Quirt, 1984). Later drusy quartz (Q3) is also found locally within the previous quartz-dissolution zones.

Illite-kaolinite-chlorite alteration halos (Fig. 9) are up to 400 m wide at the base of the sandstone, thousands of metres in strike length and several hundreds of meters vertical extent above deposits (Wasilyuk, 2002, for McArthur River; Kister et al., 2003, for Shea Creek; and Bruneton, 1993, for Cigar Lake). 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 (Hoeve and Quirt, 1987). Illite-dominated haloes have K_2O/Al_2O_3 ratios > 0.18 and MgO/Al_2O_3 ratios < 0.15; kaolinite-dominated haloes 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 < ~0.04 (Percival, 1989, from Sopuck et al., 1983). Compared to background values of 0.1-0.16 ppm 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).

The regional background of U is 1-2 ppm in lake sediments (Maurice et al.. 1981), <3 ppm in sandstone (Ibrahim and Wu, 1985) or 1 ppm in sandstone (Andrade 2002, Table 4.1, Percival, 1989; Table 4 of Wallis et al., 1985). Anomalous U (>2.5 ppm) in the above described clay alteration halos extends in places to the top of the sandstone, even in sections greater than 500 m thickness (e.g. Clark, 1987).

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 ~95 ppm. Anomalies in lake sediments reach values of 1500 ppm U (Maurice et al., 1985) in the Key Lake area. The trace elements U, Ni, As and Co are above background in the haloes above the ore deposits, but dispersion is restricted to tens of metres (Sopuck et al., 1983), thus limiting their utility as pathfinder elements.

Limited alteration is evident above "ingress-type" deposits - these are essentially "blind" in terms of exploration guides, except for geophysical methods. Many are entirely basement hosted, mono-mineralic and have very narrow, inverted alteration halos along the sides of the basement structure, grading from illite +/- sudoïte on the inside, through sudoïte +/- illite, to Fe-Mg chlorite +/- sudoïte on the outside against fresh basement rock (Figure 10; Quirt 2003). These metallurgically attractive deposits have been discovered either accidentally (the discovery drill-hole was left running "too long" through fresh basement and continued through the hanging wall of a fault into such a blind deposit) or by design of exploration geologists who understood the geometry of a fault system and suspected an "ingress-type" deposit. Some deposits have both "ingress" and "egress" characteristics (e.g. McArthur River), suggesting complex hydrothermal systems involving both processes very close to one another. A good understanding of the basement geology and structural features are requirements for "ingress -type" exploration.

Hydrothermal, ore-related alteration effects are superimposed on pre-existing alteration assemblages, including the paleo-weathering alteration recorded by the red-green profile below the basal unconformity, and the initial detrital clay mineralogy that is locally preserved in silicified zones (Mwenifumbo et al., 2005). This has led to confusion as to which process generated the red-green transition at the basement-sandstone unconformity. Ore-related white clay alteration is superimposed on the red hematitic paleo-weathering alteration (McDonald, 1980). Very dark coarse grained hematite alteration forms a cap over ore deposits (Figure 9) such as Cigar Lake (Andrade 2002); and forms dense cement within parts of the Read Formation and Bird Member of Manitou Falls Formation, particularly in units close to the basal unconformity and near U deposits. This very dense, crystalline, hydrothermal hematite alteration contrasts with the brick red, fine-grained hematite that is interpreted as a detrital or very early diagenetic mineral, and is intimately interlayered with clay minerals to form micro-laminae in oncoidal hematite beds (Yeo et al., 2005a).

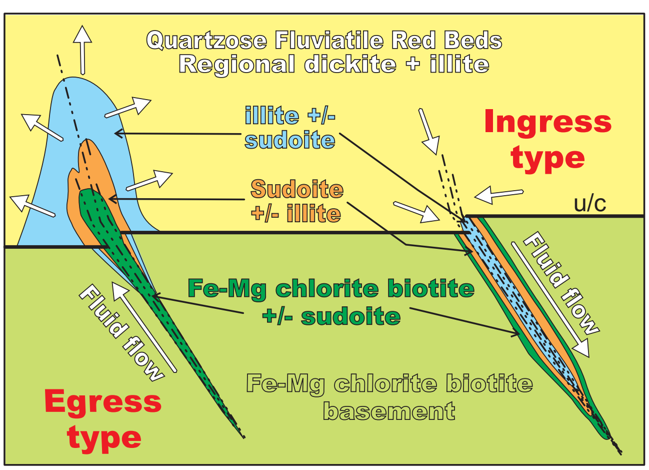

FIG. 10. Diagrammatic explanation of egress versus ingress type unconformity-associated U deposits, after Quirt (2003). U/C = unconformity. "sandstone" = siliciclastic fluviatile sequence of red sandstone, conglomerate and mudstone.

Key Exploration Criteria

Geological Exploration Criteria

A first-order exploration criterion is irregularities of the basal unconformity in Paleo- to Mesoproterozoic red-bed basins. These red-bed basins are generally flat lying, thin depressions (<2 km, see figures 4 and 5) and are filled predominantly by quartz-dominated siliciclastic sequences of fluvial conglomerate, sandstone and mudstone. Original thicknesses of these red beds were probably much greater. Other rock types are minor or cap the red bed sequences. All examples of significant unconformity-type ore identified to date are associated with such sequences, as preserved in the Athabasca Basin of Saskatchewan-Alberta, the Thelon Basin of Northwest Territories-Nunavut (Figs 1 and 2) and the Kombolgie Sub-Basin (northern McArthur Basin) of Australia. These three basins have been compared by Kyser et al. (2000) who have found that they have much in common in terms of their geological framework, but have subtle differences in diagenetic history. Several other Mesoproterozoic basins of this type are identified around the world, and do contain unconformity-U prospects, but to date no significant deposits.

Based on the Athabasca Basin example, another first-order exploration strategy for unconformity-type U deposits requires the identification of basement complexes of highly deformed and metamorphosed Archean orthogneisses and paragneisses, tectonically interleaved with Paleoproterozoic platformal sedimentary assemblages (Figure 5B). These supracrustal assemblages are characterized by relatively high U "Clarke values" and include graphitic metapelites. Late Paleoproterozoic granitoid plutons and pegmatites, generated during regional high-grade metamorphism and anatexis from the metasedimentary rocks, are rich in K-Th-U hosted by minerals such as monazite, zircon and uraninite. These have been documented both in the Wollaston-Mudjatik domains of the eastern Athabasca basement (Madore et al., 2000) and in the Taltson magmatic zone of the western Athabasca basement that has been extended past the Shea Creek area (Brouand et al., 2003) and eastward as far as the Virgin River Shear Zone by Stern et al. (2003) and Card et al. (2005a).

Second-order empirical parameters associated with unconformity mineralization include graphitic strata and fault structures within the basement complex and the presence of subtle but very significant, brittle post-sandstone structures. Ore is typically focused at the intersection of the basement-sandstone contact and high-angle oblique reverse faults that appear to be reactivated older basement structures.

These structures have propagated upward into complex splays within the sandstone. Detailed structural analysis of these splays in drill core can lead into the key basement fault zone and provide the local structural framework of a prospect. Intersections of different arrays of high-angle faults are especially significant, such as between the P2 Fault and cross faults at McArthur River (Fig. 4 of McGill et al., 1993), and between Rabbit Lake Fault trends and the Tabbernor array (LeMaitre and Belyk, oral presentation, Saskatchewan Open House, 2005).

These lithologic and structural attributes are clearly evident in the majority of unconformity-associated U deposits in the Athabasca Basin. Most of the sub-Athabasca Basin deposits are located in its eastern part, along the northeast-trending transition zone between the basement Mudjatik and Wollaston domains. This transition zone includes high proportions of metapelitic, meta-quartzitic and meta-arkosic gneiss that are isoclinally folded and interleaved with Archean granitoid gneiss. Many significant deposits in this eastern area are also located close to the ancient unconformable contact between the Archean granitoid gneiss and late Paleoproterozoic basal Wollaston Group metapelitic gneiss that contains significant graphitic units.

In the western part of the Athabasca Basin, significant deposits are also located in the basement complex exposed by the central uplift of the Carswell Impact Structure (Lainé et al. 1985), again associated with graphitic units. Although structurally complicated by the meteorite impact, these basement-hosted deposits are close to the basal Athabasca unconformity that is mainly overturned in that area (Ramaekers, 1990). High-grade intersections have been reported from a number of other localities, also associated with graphitic shear zones in the underlying basement. An Alberta example is the Maybelle River prospect (Kupsch and Catuneanu, 2005; Pana et al., 2005). Exploration and related research continue to expand the known and potential resources throughout the Athabasca Basin.

Favourable basins show geochemical evidence of large-scale fluid flow resulting in regional clay alteration (e.g. Earle and Sopuck 1989) and the development of local redox boundaries within the overall red-bed sandstone sequence. Local alteration halos of potassic clay alteration minerals (illite), boron alteration minerals (dravite), quartz cement and quartz dissolution are the main vectors for local exploration, and also form extensive corridors within which more detailed searches are conducted. These features (described above under Alteration Mineralogy and Geochemistry) are logged carefully during drill programs, with the aid of on-the-spot mineralogical analyses by short-wave infrared (SWIR) spectrometers such as PIMA II© (Integrated Spectronics Ltd.) and FieldSpec Pro, which were compared by Percival et al. (2002). Calibrated software algorithms for semi-quantitative analyses (Earle et al., 1999) enhance the usefulness of these spectrometers. Spectrometric methods have 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 (Jefferson et al., 2001).

Kyser et al. (2000) reviewed and compared the mineralogical and fluid paragenesis of the Athabasca, Thelon and Kombolgie basins in order to assess which of these parameters might be critical for the design and priority setting in exploration programs. In addition to emphasizing the great degree and extensive time involved in alteration of all three basins, they noted a number of differences between these basins that support the high prospectivity of the Athabasca Basin and downgrade that of the Thelon and Kombolgie basins. These comparisons are summarized in Figure 11, with modifications after Kyser et al. (2000) based on information summarized herein. In the EXTECH IV multidisciplinary study, mineralogy was not a strong focus but even so, data on crandallite (Mwenifumbo and Bernius, 2005) and zircon (Rainbird et al., 2005) suggest some conclusions that differ significantly from those of Kyser et al. (2000). Crandallite in the Athabasca Basin is not an early diagenetic mineral but a near-peak diagenetic mineral, closely interleaved with and enclosing clay minerals on a regional scale in the siliciclastic strata. Zircon is not altered regionally, but is overall fresh and well preserved. Minor altered zircon can be found nearly everywhere in the basin but is common in strata within alteration envelopes of ore deposits as well as in the altered part of basement. In most cases the altered zircon shows evidence of U uptake, not leaching (Hecht and Cuney, 2003 and Cuney, 2003). Severe corrosion of zircon is very local. Diagenetic xenotime that overgrows the zircon contains virtually no U, whereas the enclosing diagenetic apatite contains highly variable, locally abundant U (Rainbird, 2003b). In our view, these data form provide a framework for further work in understanding the evolution of these basins. Much more study is required to establish the regional paragenetic framework and to distinguish alteration that is a vector to ore from that which is a product of basin-wide diagenesis. Regional scale products of alteration remain useful for evaluating basin-scale mineral potential, as

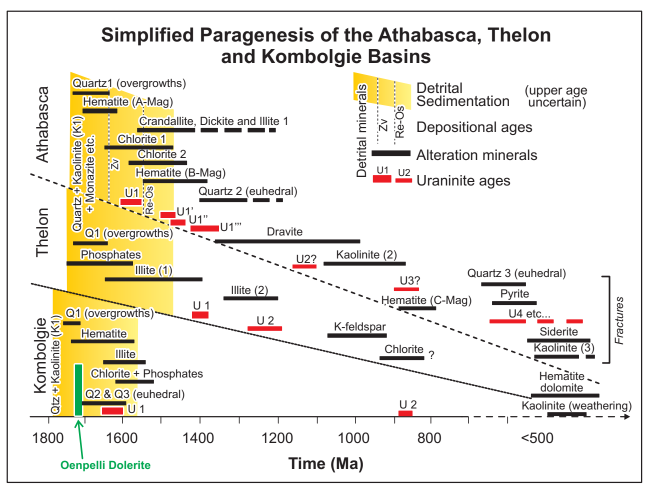

FIG. 11. Simplified mineral paragenesis of the Paleo- to Mesoproterozoic Athabasca, Thelon and Kombolgie basins, after Figure 10.21 of Kyser et al. (2000). The ages of primary uraninite, the first deposition of U in a given deposit, may be different for different deposits, hence U1 (~1500-1600) and U1' (~1350-1400) might both be primary in the Athabasca Basin. Depositional ages in the Athabasca Group are Zv: U-Pb on volcanic zircon in Wolverine Point Formation, and Os-Re: primary organic matter in Douglas Formation. Geochronology is documented and discussed further in text.

exemplified by Kyser et al. (2000).

Areas of pre-existing complexity along basement structures are particularly favourable, such as those in the Wollaston-Mudjatic domain transition zone (e.g. extensional or compressional flexures, bifurcations, splays, duplex structures and cross structures). These have undergone repeated brittle post-Athabasca Group movement as shown by compilation and interpretation of regional magnetic and EM data, detailed structural mapping and stratigraphic reconstructions (e.g. Thomas et al. 2000; Portella and Annesley 2000a, b; Ramaekers et al. 2005a, b; Yeo et al., 2005b). Specific evidence for this working model, is in detailed maps of basement rocks in the Sue 'C' open pit mine (McClean Lake) that demonstrate the association between repeatedly reactivated basement faults and brecciated, fault-hosted high grade ore spatially associated with basement graphitic schists and meta-quartzites (Tourigny et al. 2005).

Paleo-valleys on the order of 20-40 m deep developed during initial sedimentation of the overlying Athabasca Group, in spatial and temporal association with the above fault zones. The concept of paleo-valleys spatially associated with Athabasca ore deposits was introduced by Wallis et al. (1985) if not earlier, but little further had been published on this topic until detailed structural, stratigraphic and sedimentologic studies in the Dielmann Pit by Harvey and Bethune, 2005; Collier and Yeo, 2000; and stratigraphic-sedimentary studies in Sue C Pit by Long, 2001, 2005). Detailed stratigraphic analysis of drill cores at the McArthur River (Bernier, 2004) and Wheeler River areas (Jefferson et al. 2001, Yeo et al. 2005b) has shown that basement faults were repeatedly reactivated in alternating reverse and normal offsets throughout deposition of the preserved portion of the Manitou Falls Formation, creating lateral thickness and facies changes, thereby hampering correlations. Ramaekers and Catuneanu (2004) have put these local observations into regional stratigraphic context. Their analysis of basin development outlines areas of greatest basement flexure during and after sedimentation. Concomitant brittle reactivation of basement structures through time improved the permeability of basement structures in these areas of flexure as conduits for mineralizing fluids. The combined association of these geological features provides strong encouragement for further exploration near such basement flexures.

Geochemical Exploration Criteria

Lake water and sediment geochemistry (e.g. Coker and Dunn, 1983; Maurice et al., 1985) and radiometric prospecting (Reeves and Beck, 1982) were significant tools in early regional exploration. Another early geochemical exploration technique involved measuring and contouring radon gas emission as an expression of radioactive decay related to underlying U ore deposits on reconnaissance to detailed scales (e.g. Dyck, 1969; Scott, 1983). 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 trees sampling anomalous ground water (Dunn, 1983). Groundwater itself is a useful exploration medium (e.g. Toulhoat and Beaucaire, 1993), especially given the long history of fluid flow and the still-active but variably constrained groundwater systems in the broadly permeable Athabasca Group, and tightly constrained potential aquifers in faulted basement rocks (Cramer, 1986).

As exploration advanced to deposits at greater depths, focus shifted to the above-noted alteration mineralogy, both regional and local, that is reflected by surficial geochemistry. Large halos of potassic clay alteration minerals (illite), boron alteration minerals (dravite), quartz cement and quartz dissolution are intersected in various places at the present surface (e.g. Shives et al. 2000), 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., 2005) and by gamma ray spectrometry as outlined below.

Gamma ray spectrometry is here treated as a geochemical tool, because it directly measures U, K and Th in surficial material. Campbell et al. (2005) 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 (Campbell et al., 2002) as geochemical prospecting and lithologic mapping tools. Interpretation of results from such surveys requires knowledge of paleo-ice-flow directions and till stratigraphy. The advancement of data manipulation and presentation as ternary ratios of K, Th and U on a single map makes it easier to interpret inter-relationships of these elements, and to define trends that may not be evident on simpler presentations of exactly the same data (e.g. Campbell et al., 2002 compared to Richardson, 1983 for the same area). The extensive illite alteration corridor between McArthur River and Key Lake (Fig. 6) 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. A detailed airborne gamma ray survey is recommended to test this proposed correlation.

Quaternary deposits are strong indicators of local bedrock and a variety of ice flow directions must be considered in tracing surficial materials back to their sources (Campbell, 2005). In the Athabasca Basin, this bedrock is broadly the basement gneiss or the Athabasca Group 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. 2005). 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. Detailed follow-up exploration has traditionally focused on airborne and ground electromagnetic methods (e.g. Matthews et al., 1997) which have been and remain the most effective tool to identify the precise location, depth and characteristics of basement conductors that correlate with graphitic shear zones, along which ore deposits are commonly located.

Electromagnetic methods also can be used to detect ore-related alteration features. A high-resolution airborne electromagnetic technology, termed Tempest, has been used in Australia to detect shallow but hidden, low-resistivity alteration zones, and to crudely map fault offsets of the unconformity (Bisset, 2003). Recent testing of improved audio-magnetotelluric methods in the McArthur River area of Athabasca Basin has demonstrated the ability to detect both deep conductors and alteration zones (Craven et al. 2005). Highly altered, clay-rich, quartz-corroded quartzarenite has relatively low resistivity, whereas quartz-rich silicified zones are characterized by high resistivity (Craven et al. 2005). 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, 2005).

Airborne magnetic surveys provide the means to map basement geology from the margins of these Proterozoic basins to their centres (e.g. Figure 1 of Fogwill, 1985; Pilkington 1989), with the aid of magnetic susceptibility and related data from outcrop and drill core that intersects the basement. Thomas and McHardy (2005) provide a modern review of this technology and demonstrate its application to the eastern Athabasca Basin, pointing 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 Group. Exploration geologists use such basement features to focus grass-roots programs and as a measure of prospectivity to help set priorities within claim groups, e.g. Moore Lakes Prospect of JNR Resources Inc. (# 21 of Fig. 4 and Table 1), P-Patch just east of Key Lake (Madore et al., 2000).

Seismic reflection is a relatively new exploration tool from the mineral industry perspective. Its main contribution has been to provide continuous structural framework transects in two, and locally in three dimensions (White et al., 2005). These transects can focus on either shallow (surface to a few km below the unconformity, e.g., Gyorfi et al., 2005) or deep (unconformity to Moho, e.g., Hajnal et al., 2005) framework geology questions by varying frequency, spacing and data processing. Sonic and other rock quality data from drill holes located as close as possible to the seismic lines are calibrated with the aid of borehole geophysics (Mwenifumbo et al., 2004). Seismic data offer the only way of explicitly imaging laterally continuous detailed structural features, from which complete structural sections can be interpreted using modern structural geological analogues, such as comparative data from mapped outcrop geology located on strike, outside the basin (e.g., Tran, 2001; Gyorfi et al., 2005). Fundamental exploration parameters such as the location and irregularities in the unconformity, and shallow to deep faults can be profiled in this way.

Gravity transects (or airborne gravity) can detect alteration zones as negative gravity anomalies (de-silicified 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, 2005). Gravity also provides insights into the geological framework on both regional and district scales.

Genetic / Exploration Models

Conventional Models

Why are such high-grade, large-tonnage U deposits found only at the basal unconformities of shallow, late Paleoproterozoic to Mesoproterozoic siliciclastic basins? And why is the Athabasca Basin so special with deposits one or two orders of magnitude larger than similar deposits elsewhere in the world? Conventional models for unconformity-associated U deposits invoke late diagenetic to hydrothermal processes. Most models in use today are a combination of empirical spatially associated attributes invoking diagenetic-hydrothermal processes and ore formation being spatially and temporally focused by the reactivation of pre-Athabasca Group structures (e.g. Hoeve and Sibbald 1978; Hoeve et al. 1980; Kotzer and Kyser 1995; Fayek and Kyser 1997; Tourigny et al. 2005).

These models propose that oxidizing U-transporting basin fluids, heated by geothermal gradient, eventually attained 200°C (~ 5-6 km) at the unconformity and reacted with basement graphite to create methane (CH_4) prompting U precipitation due to mixing of reduced and oxidized fluids (Hoeve and Sibbald, 1978). Precipitation was primarily focused by structural and physiochemical traps (Thomas et al., 2000). These traps operated in fixed locations for very long periods of time (Hoeve and Quirt, 1987), perhaps hundreds of millions of years. Zones of fluid mixing are characterized by alteration halos that contain illite, kaolinite, dravite, chlorite, euhedral quartz, and locally, Ni-Co-As-Cu sulfides (Kotzer and Kyser, 1995). The latter described the chlorite as Mg-chlorite (=clinochlore) but it is actually a less common Al-Mg-chlorite termed sudoïte (Percival and Kodama, 1989).

In the above models, fluids became mixed where reducing basement fluids circulated upward into the overlying oxidized formational-fluid environment (egress type). Ingress of basin formational fluids downward into the basement developed inverted alteration zonation, mainly in host basement rocks, and has virtually no expression in the overlying siliciclastic strata (Quirt, 2003). Quirt and Ramaekers (2002) and Quirt (2003) have reminded us that there are many variations on the ingress and egress themes in the unconformity-associated U deposit model.

Due to the geological and mineralogical variations in deposits of the Athabasca Basin, many variations on this concept have been proposed, as summarized by Tremblay (1982) and Ruzicka (1996a). Two other fundamentally different models (summarized by Hoeve et al., 1980) have been proposed and, while not coherent with current knowledge, are consistent with parts of the above model. Knipping (1974), Langford (1978) and to some degree Dahlkamp (1978) proposed a mainly supergene origin for these deposits related to pre-Athabasca Group weathering of basement rocks, transport by surface and ground waters, and deposition in basement host rocks under reducing conditions before the Athabasca Group had covered them. In the 1970's, deeply buried deposits were not known, but it is now very clear that these deposits were formed after at least the lower Athabasca Group was deposited, because the extent of U and alteration minerals is throughout the local stratigraphic column. A magmatic hydrothermal origin was briefly considered (summarized in Hoeve et al., 1980) but there is no evidence of magmatism coeval with U deposition, the closest magmatism being sparse tuffs in the Wolverine Point Formation that are dated as 1644±13Ma (U-Pb) on volcanic zircon by Rayner et al. (2003). The source of these tuffs is interpreted as being very distal to the basin (Rainbird et al., 2005) and they are at least 50 Ma older than all uraninite ages, the oldest being some Pb-Pb results between 1550 and 1600 Ma by Alexandre et al. (2003).

Advances of the Last Decade in Genetic Models

The various conventional models all try to account for the combined efficiency of source, transport and deposition of U, as summarized by Cuney et al. (2003). With more than 35 years of research after discovery of Rabbit Lake, advances have been significant, many new questions have arisen, and some of the fundamental enigmas of Hoeve and Sibbald (1978) remain. Like lode Au deposits (Poulsen, 1996) a very wide variety of models exist for U deposits (Cuney et al., 2003): "U-deposits may appear at each step of the geological cycle, from magmatic and fluid fractionation in the deep continental crust (such as the Tranomaro pyroxenites, Madagascar; Rössing alaskites, Namibia: 750-800°C, 5-7 kbar) to evapotranspiration at the surface (such as the Yeleerie calcretes, Australia). However, very high-grade, large-tonnage U deposits are only found as the unconformity-associated type, of Mesoproterozoic age. In the following reviews of source, transport and deposition of U, we suggest that there is sufficient diversity in unconformity-associated deposits to also require multiple models, or variants of a main model (e.g., Quirt, 2003).

<u>Uranium Sources</u>

The above models involve a number of possible U sources, ranging from primary first-stage U concentration through to the immediate sources of the ore-forming solutions. Tremblay (1982) proposed that a combination of events and sources through time were required to ultimately result in a world-class deposit. Others have focused on the penultimate reservoir of U, considering that the various primary sources were not important because of the efficiency of U transportation and focusing mechanisms that created the ore deposits.

Primary sources of U include radiogenic S-type granites and pegmatites (e.g. Madore et al., 2000), metasedimentary terrains with abundant pelite whose U endowment is well above Clarke values (Miller and LeCheminant, 1985), and previous uranium concentrations such as in the Wollaston Supergroup (Delaney, 1993; Yeo and Delaney, 2005), pegmatites (Thomas, 1983) that intrude the Hearne craton (formerly Cree Lake Zone) and the Beaverlodge camp (Koeppel, 1967, Tremblay, 1972; Ruzicka, 1996b). As in Cu provinces of the world, regions relatively well endowed with U (such as the Wollaston and Mudjatik domains of the Trans Hudson orogen), 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 (in particular the Wollaston and Mudjatik domains) and western Trans-Hudson orogen (e.g. Peter Lake domain and Wathaman Batholith). These basement domains were the main sources of primary sediment and fluid for the Read and Manitou Falls formations that overlie most deposits of the eastern Athabasca Basin.

Madore et al. (2000) documented that monazite from leucogranites in the Wollaston-Mudjatik domain contain abundant U, and this would have been in addition to the many known uraninite showings and uraninite disseminated in pegmatite. They calculated that sufficient U could have been directly extracted from monazite in a hypothetical volume of such rock in a shear zone 25 km long x 1 km wide x 5 km deep (a volume of 1.25 x 107 cc) to have formed the McArthur River deposit (in the order of 250 kt U). On the other hand, Ramaekers et al. (2005a, b) have shown that the third stratigraphic sequence and components of the first and second sequences of the Athabasca Group were derived from the south, parts of the second sequence were derived from the north, and the late diagenetic history of the Athabasca Basin involved fluid flow from the west, therefore the full history of the various sources of U for the basin, and how U was transported by basin fluids is much larger and complex. A full assessment of this question is beyond the scope of this paper, except to note that the other two basins hosting significant unconformity U deposits also overlie such terrains and have similar metallogenic histories, namely the Thelon Basin of Nunavut (Miller and LeCheminant, 1985; Kyser et al., 2000) and the McArthur Basin of Australia (Kyser et al., 2000).

Intermediate-stage (or penultimate) reservoirs of U were the fluids and the sediments that came from the above ultimate sources. A number of workers from Macdonald (1980) to Ruzicka (1996) have held that the sediments were the main reservoirs for the metals. Some models reject the basin-filling strata as a significant source of U, noting that minerals capable of yielding U are absent (Tremblay 1982, Cuney et al., 2003), or focus on the relatively high U contents in unaltered local basement rock (Annesley et al. 1997; Hecht and Cuney, 2000; Madore et al., 2000). There is no doubt that hydrothermal alteration of basement rocks preferentially released U, therefore this must have been one source of U for the deposits. Nevertheless, the question of mass balance must be considered to assess the relative importance of potential U sources. .

The significance of basement as a direct U source is constrained by three key factors: permeability, surface area over which alteration reactions could take place, and the volume of basement rock affected by such alteration. Permeability in fresh basement rocks was very low except in local fracture zones. Surface area for chemical reaction was limited, by lack of permeability, to the area of the specific alteration front. The documented volume of basement affected by alteration is limited to the area of the Athabasca Basin times the thickness of altered basement: 10-50 m. Much deeper alteration can be observed along some fault zones but may be attributed to paleo-weathering. Basement drilling clearly documents sharp transitions to unaltered, very tight rocks outside of shear zones (see Fig. 10). In contrast, the siliciclastic strata filling the Athabasca Basin had high permeability, and a conservative thickness of 1 km below aquitards such as shale in the Wolverine Point and Douglas formations. Due to the high permeability, the entire surface area of each sedimentary grain was available for chemical reaction, with minor restrictions such as matrix clay and local pre-ore silicification. The relatively high abundance of matrix clay in the Fair Point Formation raises a question of its permeability, but the degree to which Cretaceous oil infiltrated this unit suggests that permeability was sufficient for considerable fluid flow. This provides the sedimentary basin fill several orders of magnitude more significance as a source of U, demanding a re-assessment of its primary U content.

Cuney et al. (2003) stated that "the average U-content of the sandstones, away from mineralized areas, is below 1 ppm, and 50 to 80% of U is now within zircon". Zircon was effectively altered during diagenesis, but altered zones were enriched in U together with Ca, LREE, Al, and P (Hecht and Cuney, 2000 & 2003). Mathieu et al. (2000) also noted that detrital monazite may have been a source of U, because of its alteration during diagenesis to Ca, Sr, LREE Al-phosphates at the basin scale as observed in the Franceville basin; they also considered that monazite alteration was the major source of U for the Oklo area deposits. Cuney et al. (2003) noted that the average Th content of the sandstone is below 10 ppm, except in the lower Manitou Falls Formation of the eastern Athabasca Basin. From this he deduced that the amount of U derived from monazite was limited. He considered that part of any liberated U was trapped in altered zircon, in Ti-oxides and hematite deriving from detrital Fe-Ti oxide alteration. Last, but not least, he considered that the amount of U which may have been present as U-oxide must have been very low owing to the highly oxidized conditions that prevailed during deposition of mid-Proterozoic, mainly continental, sandstone that was devoid of organic matter.

Mwenifumbo and Bernius (2005) corroborated the low average U content and provided an alternative interpretation. Although Cuney interpreted the low U content as showing that the sandstone was a poor source of U, Mwenifumbo and Bernius (2005) and some authors of this paper here submit this as evidence that U was preferentially removed from the siliciclastic strata during peak diagenesis, as part of the mineralization process. In addition, Ibrahim and Wu (1985) noted regional background levels of 40 ppm Th in the area of the Midwest deposit, thus calibrating the regional airborne gamma ray anomaly in Th over the eastern Athabasca Basin, and the potential for original high monazite. The removal of U is also qualitatively deduced because the source domains from which these detrital strata must have been derived (e.g. Wollaston and Mudjatik domains, Wathaman batholith) contain above-Clarke abundances of U as shown by Madore et al. (2000).

Even if only a very small fraction of U was liberated from the sediment during diagenesis (much smaller than the proportion considered by Madore et al., 2000), the huge volumes would have easily supplied sufficient U to create the major deposits. No enrichment was required other than that carried by the detrital grains from the source basement rocks. The key question is - when was the U removed from the detritus? Petrographic analysis suggests that at least some of this removal took place during the extensive burial diagenesis that pervasively affected the entire Athabasca Group, because lithogeochemical analysis by Quirt (tables V and VI in Sibbald et al. 1976) suggests these rocks were originally arkosic, and this suggestion has been supported by textural studies of Bernier (2004), Collier (2003), Kupsch (2003), Kupsch and Catuneanu (2005), Ramaekers (1990) and Ramaekers et al. (2005b).

Because virtually no feldspar is actually preserved mineralogically, the above interpretation hinges on geochemical and textural evidence (i.e. clay-filled spaces in the detrital framework that might have been feldspar or other labile grains), and cannot be proven with available data. Macdonald's (1980) opinion was that paleo-weathering was so extreme that no detrital feldspar survived transport into the basin. Cuney (pers. comm. 2005) has noted that regional illite is developed only in certain corridors where basement structures crosscut the basin, such as the McArthur River - Key Lake corridor (Fig. 6) and at Shea Creek (Rippert et al., 2000). Where no basement structure is observed, as in the Erica 1 or Rumple Lake drill holes, most of the primary kaolinite is preserved, and illite is poorly developed. The question arises, if there was sufficient K available in the whole basin from K-feldspar alteration, should illite have been developed everywhere, or could K have moved from depleted areas into the illite zones?

Consideration of the detrital heavy mineral component also suggests bulk removal of U during peak diagenesis. Studies of crandallite group, viz., aluminum phosphate (AP) minerals found within the Manitou Falls Formation (Mwenifumbo and Bernius 2005) demonstrate that these have high complements of Th and REE but are very low in U. Although no detrital monazite is preserved in the sandstone, the AP are very similar to AP alteration products of monazite that have been documented in basement rocks underlying the Athabasca Group (Cuney et al. 2003, Hecht and Cuney 2000; Madore et al. 2000). Similar alteration products have been noted in both basement and alteration zones proximal to Australian U deposits in the Kombolgie Basin (Gaboreau et al. 2003). The Australian AP minerals have the same relative proportions of light REE as the associated monazite, which they replace. In the Athabasca Basin, AP minerals are concentrated as diagenetic cement in coarse-grained beds of the Manitou Falls Formation. These AP minerals were described by Kyser et al. (2000) as "early diagenetic" crandallite but were not included in their charts. The AP minerals analysed by Mwenifumbo and Bernius (2005) are also diagnosed as crandallite, but are intimately intergrown with clay minerals, hematite (H1?) and chlorite that form a mid-diagenetic or early phase of a moderately high-temperature (150-170°C) diagenetic assemblage which post-dates Q1 quartz overgrowths (Figure 11).

Conglomeratic sub-units in the Bird Member of the Manitou Falls Formation (Table 4) commonly include sandstone interbeds whose horizontal and trough cross bedding,

Table 4. Summary of lithostratigraphic units and unconformity-bounded sequences of the Athabasca Group (after Ramaekers,1990).

Formation [code] This study	**Member**; this study [code] (textural lithology) MTG = maximum transported grain size in mm [s] = present only Saskatchewan	**Formation** (Ramaekers 1981 & earlier)	**Inf. members** (Ramaekers 1990)	**Sequence & Deposystem** This study
Carswell [C]	Upper and lower carbonates (dololutite, dolorudite, stromatolite, oolite, dolarenite) [s]	Carswell	CF	4
Douglas [D]	(dark grey organic-rich mudstone with desiccation or synaeresis cracks and interbeds of fine to very fine quartzarenite, MTG <2) [s]	Douglas	DF	
	Now recognized as up-faulted Locker Lake Formation [s]	Tuma Lake	Unrecognized	
Otherside [O]	Birkbeck [Ob] (quartzarenite with minor thin interbeds of dark mudstone near the top; MTG <2 except for pebbly unit near base)	Otherside	OF	
	Archibald [Oa] (quartz-pebbly quartzarenite, quartzarenite; MTG<8)			
Locker Lake [LL]	Marsin [LLm] (quartz-pebbly quartzarenite; MTG 8-16)	Locker Lake	LL	
	Brudell [LLb] (thin-conglomeratic quartzarenite; MTG >16)			
	Snare [LLs] (quartz-pebbly quartzarenite; MTG 2-16, sparse mudstone <50 cm)			
	Unconformity	Unrecognized	Local unconf?	Unconformity
Wolverine Point [W]	Claussen [Wc] (interstitial-clay-rich quartzarenite, sparse mudstone interbeds <1m thick; MTG <2)	Wolverine Point	WPb	3
	Brule [Wb] (interbedded mudstone >50 cm and tuffaceous quartzarenite, common thin intraclast conglomerate; MTG <2 except for local basal lag			
Lazenby Lake [LZ]	Dowler [LZd][2] (quartzarenite, minor siltstone and quartz-pebbly quartzarenite; MTG < 8)		WPa	
	Larter [LZl] (quartz-pebbly quartzarenite, minor mudstone intraclasts; MTG <8)			
	Shiels [LZs] (quartz-pebbly quartzarenite with pebbly layers, rare mudstone beds and intraclasts; MTG >8)			
	Clampitt [LZc] (pebbly base, quartz-pebbly quartzarenite, minor laminated siltstone & mudstone; MTG <8)			
	Hodge [LZh] (5-30 cm basal conglomerate, quartz-pebbly quartzarenite and conglomerate, sandstone intraclasts; MTG >8)	Lazenby Lake	LzL	
	Basal unconformity to Mirror Basin	Unconformity	Unconformity	Unconformity
Manitou Falls [MF]	Dunlop [MFd] (>1% clay-intraclasts in quartzarenite, mudstone interbeds; MTG <2)[s]	Manitou Falls [MF]	MFd, part MFc	2
	Collins [MFc] (quartzarenite with minor quartz pebbly beds, mudstone interbeds, <1% clay intraclasts, <2% conglomerate interbeds)		MFc	
	Warnes [MFw] (quartzarenite and clay-intraclast-rich quartzarenite in Karras Deposystem, from Virgin River area to Alberta)		Original MFa	
	Raibl [MFr] (quartz-pebbly quartzarenite in Moosonees Deposystem, northeastern Athabasca Basin; minor clay intraclasts, <2% quartz-pebble conglomerate; MTG >2)[s]		MFd (part) MFc (part)	
	Bird [MFb] (interbedded >2% quartz-pebble conglomerate, quartz-pebbly quartzarenite, thin mudstone & siltstone interbeds; MTG >2)		MFb	
F-O: Undivided Fair Point to Otherside formations in Carswell Structure[s]			William River S...	
	Local unconformity separates Manitou Falls and Read formations[s]		Not recognized	
S/M: undivided Smart or Manitou Falls formations (only in Alberta)			MFc	
Smart [S]	(quartzarenite with local red mudstone and oncoid interbeds at base). May be a distal equivalent of Read Formation		MFc	
Read [RD]	(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)[s]		MFa in Industry drill logs	
Reilly [RY]	RYcg (conglomeratic quartzarenite)[s]		MFb	
	Basal Unconformity to Reilly Basin[s]	Basal unconformity		Unconformity
Fair Point [FP]	Beartooth [FPb] (0-10% quartz»lithic-pebbly quartzarenite with abundant matrix clay; MTG generally <64)	Fair Point	FP	1
	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			
No formal designation	BL (basal lag, pebbles to boulders)	Basal lag	Basal lag	

Blue shading indicates formations and members that have been renamed, formalized or otherwise revised from Ramaekers (1990). The riginal map codes are preserved for continuity except where the lithology is significantly changed. The framework mineral for all textural types from conglomerate to mudstone is 99% quartz. Every unit contains cross bedding and ripple cross bedding, and most contain 1-layer thick quartz pebble or granule beds. Only diagnostic stratigraphic parameters such as grain-size and desiccation cracks are specified here.

and sieve textures are finely outlined by black laminae that strongly resemble black sand paleo-placer segregations (e.g. Figure 5 of Yeo et al., 2000). In thin section these are seen to be mainly cement of dense hematite and AP minerals surrounding quartz grains, with zircon as the only commonly preserved heavy mineral (Mwenifumbo and Bernius, 2005). The sand-grain-size, sedimentary structures and inferred hydrodynamic characteristics of these black laminae are consistent with them being relict heavy mineral layers, with the heavy minerals now altered to a mixture of AP, specular hematite and chlorite as cement to the remaining detrital grains of quartz and zircon. Uranium could also have been incorporated in the Athabasca basin-fill by adsorption to detrital clay minerals and iron oxide weathering products that served as chemical sponges during physical erosion and transportation (see Macdonald, 1980). We therefore conclude that the Athabasca Group formed a repository of detrital material with volumetrically significant primary U, and from which one or more post-sedimentary processes removed U. This hypothesis can be tested by lithogeochemical analysis of bulk samples of the Athabasca Group and of basement source terranes, followed by quantitative modelling of the likely original sediment composition in comparison to the present composition (Experimental work could assess what ppm level of U could be adsorbed under various atmospheric conditions). This is directly analogous to mass-balance calculations for the Coates Lake sedimentary basin that was inferred to have served as the penultimate reservoir of Cu, quantitatively transported during late diagenesis, by formational fluids, to form the Redstone Cu deposit (Lustwerk and Wasserman 1989). A similar exploration tool is geochemical modelling of mafic-ultramafic intrusions that uses depleted nickel and platinum values to infer their segregation into magmatic ore deposits (e.g. Fedorenko 1994).

Zircon in the Athabasca Basin was variably altered during diagenesis. Kyser et al. (2000) showed that significant zircon was destroyed during peak diagenesis in some samples from the Athabasca Basin, but did not observe this in the Thelon basin. They inferred that alteration of zircon released sufficient U to form major ore deposits in the Athabasca Basin but not in the Thelon. On the other hand, Hecht and Cuney (2000, 2003) and Brouand et al. (2003) have determined that altered zones in many zircons from sites distributed across the Athabasca Basin and in the basement are actually enriched in U together with Ca, LREE, al., P. They also contended that some of the U released by destruction of primary minerals could have been conserved by adsorption to iron oxide and clay minerals. A third perspective is that detrital zircon grains sampled regionally through the Athabasca Basin are well preserved, suitable for U-Pb dating, no more altered than other detrital zircons around the world, and have normal U contents (Rainbird et al. 2005; Rayner et al., 2003). Percival (1989) found that zircon in the Cigar Lake alteration halo do not have alteration rims, yet the alteration halo in bulk contains up to 1000 x chondrite values for REE. The common preservation of zircon yet near complete lack of most other heavy minerals on a regional basis could be taken to mean that the samples having the above-noted enrichments (Hecht and Cuney, 2000) may actually be within certain parts of the alteration zones of U deposits and therefore reflect unique mineralization processes rather than U sources. On the other hand, altered zircon crystals are generally removed during the preparation of samples used for geochronology because the altered part of these zircon is generally metamict and thus rejected without further study. Overgrowths of xenotime on regional samples of zircon, spanning the breadth and thickness of the Athabasca Basin, well away from ore deposits, contain anomalously low U (Rainbird, 2003b). Their U contents are too low to permit U-Pb dating. If these grains could be dated, they could constrain at what time U had been removed from the system, perhaps setting an upper limit on ore-formation.

The concept of zircon being the sole detrital heavy mineral as a source for U is inconsistent with the range of heavy minerals that is known (e.g. Annesley et al., 1997) in the basement provenance terranes. Therefore the sole preservation of zircon also supports the hypothesis that the other heavy minerals were preferentially destroyed either at source (e.g. pyroxene, amphibole and uraninite) or during diagenesis (e.g., monazite, fluorapatite and magnetite).

Detrital monazite would have been a significant U source, because of its abundance in basement rocks and its high U content there (Madore et al., 2000). Alteration of monazite during diagenesis to Ca-Sr-LREE-Al-phosphates at the basin scale was observed in the Franceville basin (Cuney and Mathieu, 2001), and its loss of U during incipient alteration is documented in Wollaston domain basement rocks (Madore et al., 2000; Cuney et al., 2003). The average Th content of the Athabasca Group is below 10 ppm in samples analysed by Cuney et al. (2003), suggesting to them that the amount of U derived from monazite was limited. On the contrary, the lower Manitou Falls Formation of the eastern Athabasca Basin has high Th (e.g. 30 to 730 ppm), in coarse-grained beds. The same Th-rich beds have relatively low K and U. Selected analyses of the upper Manitou Falls Formation and the underlying Read Formation corroborate low Th (1.5-6 ppm) (Mwenifumbo and Bernius 2005), like the results of Cuney et al. (2003). Therefore the lower Manitou Falls Formation is clearly anomalous.

Gamma Ray drill logs throughout the eastern Athabasca Basin (Mwenifumbo et al., 2005; Yeo et al., 2005b) show similar and very consistent trends of high Th in the lower Manitou Falls Formation. Regional airborne gamma ray data (NATGAM database of the Geological Survey of Canada) also demonstrate that the lower Manitou Falls Formation stands out as an anomalous Th-high, U-K-low trend wrapping around the entire eastern end of Athabasca Basin. Cuney et al. (2003) theorized that if part of the liberated U was trapped in altered zircon, in Ti-oxides and hematite derived from detrital Fe-Ti oxide alteration, the bulk transfer of U might have been small, but on the contrary the same gamma ray drill logs and airborne gamma ray data document quantitative, bulk, regionally low U contents. Using the same mass-balance approach as Madore et al. (2000) for U depletion from monazite in basement shear zones, and assuming that original U contents were similar in detrital monazite, 0.1 % modal monazite would have contained 4.2-12.6 ppm U in rock. With U contents now <1 ppm, and assuming 25% remobilization and precipitation, the amount of U derived from the volume of MFb and MFc (>200 km x 30 km x 0.2 km = 1200 km^3) would have been an order of magnitude larger than the 250,000,000 kg U (the size of

McArthur River) calculated by Madore et al. (2000) from a 125 km³ shear zone with the same assumptions. The Fair Point Formation can be considered as a relatively immature sedimentary source of U in the western part of the Athabasca Basin. Gamma ray spectrometric data are required to investigate this hypothesis. The point here is to consider the potential of both basement and sedimentary sources of U in genetic models.

The Read Formation (formerly MFa) underlies Manitou Falls Formation in the eastern part of the Athabasca Basin, and contrasts strongly in its very low and erratic contents of Th (as well as K and U), abundant red mudstone with desiccation cracks (lacking in Manitou Falls Formation), and low-angle cross bedding (Ramaekers et al., 2005b). The Read Formation contains abundant evidence of highly oxidizing conditions during sedimentation and this is the formation that immediately overlies the deep red regolith. Clearly the amounts of monazite and other labile minerals that contributed to its primary sedimentation were small. Evidence from the Read Formation (but not Manitou Falls Formation) is consistent with the statement by Cuney et al. (2003) that the amount of U which may have been present as U-oxide (in the original sediments) must have been very low owing to the highly oxidized conditions prevailing during its deposition, devoid of organic matter. Thus the Read Formation is not considered to have been a good source of U.

Transport of Uranium

Here we consider transport of U from its penultimate source to the site of ore deposition in aqueous solution. Factors constraining the development, movement and mineral chemical changes accompanying fluid movements in sedimentary basins are treated extensively by a series of papers in Kyser (2000), culminating in a comprehensive analysis of the three best-known Paleo- and Mesoproterozoic sedimentary basins that host unconformity-associated U deposits (Kyser et al., 2000). They have documented that the histories of fluid movements in these basins lasted 100's of millions of years at high temperatures, and low-temperature fluids are still in circulation. They compared and contrasted early, peak and late diagenesis in the Athabasca, Thelon and Kombolgie basins, developed from the relationships between mineral paragenesis and the fluids that would have been in equilibrium with those minerals. Most authors agree that 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 inferred by Cuney et al. (2003) based on the lack of organic matter in the sandstone. This is supported by the pervasive red hematitic Liesegang banding. Recent oxidation is documented by limonitic alteration along fault zones.

The passage of late diagenetic reducing fluids (part of the ore-deposition process?) is also recorded by the reduced margins of red mudstone beds and intraclasts in the Read Formation that immediately overlies the McArthur River deposit, by the drab grey and tan mudstones that characterize the Manitou Falls formation and were never fully oxidized (Ramaekers et al. 2005b), and by the presence of hydrocarbons and bitumens surrounding uraninite in many of the ore deposits. Proposed origins of these hydrocarbons range from the migration of hydrocarbons through the basin at least twice (Wilson et al., 2005) to abiotic synthesis (e.g. McCready et al., 1999 and 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 was being altered during diagenesis to form these mineralogically highly mature quartzose siliciclastic rocks. Early diagenetic brines preserved in detrital quartz overgrowths are NaCl-rich and inferred by Cuney et al. (2003) to have been derived from evaporitic layers that once existed in the upper part 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 basement Ca-rich rock types. Cuney et al. (2003) consider that high-Ca content 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., 2000). It is here considered that these same conditions would have applied to alteration of detrital monazite in the Manitou Falls Formation, as proposed by Mwenifumbo and Bernius (2005) for the Athabasca Basin and Cuney et al. (2003) for the Franceville Basin.

Cuney et al. (2003) stated that no experimental data exist to quantify the effect of such fluid compositions on U-solubility, that U solubilities of 30 ppm were calculated by Raffensperger & Garven (1995) for 5 molal Na-Ca-Cl solutions at 200°C for a fO_2 of -20 well within the hematite field, and that the concentrations of other possible strong U-ligands (e.g., F and P) is only limited by the solubility product of fluorite and apatite. Discussion of such geochemical constraints is beyond the scope of this paper.

Temperature regimes within the Athabasca Basin during the primary mineralization event are interpreted to have been in the order of 180-250°C (Pagel et al. 1980; Kyser et al. 2000, Cuney et al. 2003), suggesting gradients in the order of 35°C/km. Ramaekers et al. (2005a) suggest that either the geothermal gradient beneath the Athabasca Basin was anomalously high (40-50°C/km for a 5 km thick basin-fill) or the basin-fill was much thicker before erosion. Fluid inclusion studies by Derome et al. (2003a) indicate that T and P close to the unconformity decrease from the "early diagenetic" (160-220°C and 1 to 1.25 kbar respectively from Rabbit Lake and Carswell deposits) to the mineralization stage (140-160°C, 0.6 kbar). In contrast Kyser et al. (2000) interpreted that mineralization took place during peak diagenesis (180-240°C). Derome et al. (2003b) found that a late, low saline, CH4-bearing, higher T 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). The generally accepted protracted fluid history in the Athabasca Basin, the wide range in uraninite ages (below), and the regional alteration of the Athabasca Basin now challenge researchers to tackle regional background samples (e.g. Pagel, 1975) to help place these ore forming fluid events in a basin-wide framework.

Focus of Uranium Deposition

Structural controls on uraninite deposition have the most direct impact on exploration and development of advanced prospects. The above-noted intersections of reactivated basement shear zones with offsets of the unconformity, intersections of different steeply dipping fault arrays have directly focused uraninite deposition and now guide mine-scale exploration and development (e.g. LeMaitre and Belyk, oral presentation on new discoveries at Eagle Point, Saskatoon, November 2004). A series of studies by Baudemont and various co-authors have provided insights to fluid flow and ore locations at McClean Lake (Baudemont and Paquet, 1996), at Cluff Lake (Baudemont and Federovich, 1996) and at Cigar Lake (Baudemont and Rafini, 2000). Similar work during active mining at Sue C Pit, McClean Lake (Tourigny et al., 2002, 2005) has shown that the geometry of individual ore lenses and pods together with structural elements in the enclosing shear zone can predict the overall geometry of the deposit. He has suggested that en-echelon arrays of uraninite veins at Sue C Pit may represent mineralized hybrid extensional-shear fractures (cf. Cox et al. 2001).

Knowledge Gaps Of Thematic And Applied Research At The District And Deposit Scales

As indicated above, considerable recent progress has been made regarding source analysis, fluid compositions, flow paths and temperature regimes, but more work remains to constrain this very complex story. The hydrodynamic regime of fluid circulation is similarly not well constrained. Cuney et al. (2003) stated that brines percolated not only at the base of the basin and in the regolith, but also deeply into the basement as indicated by U-deposits rooted down to 400m into the Athabasca basement (Eagle Point) and more than 1000m in the Kombolgie Basin (Jabiluka). They also documented preserved brines in fluid inclusions from samples collected several hundred meters below the unconformity of both Athabasca and Kombolgie basins. Hectometre-scale convection cells are suggested by the 3D modelling of clay alteration halos around the deposits (Bruneton, 1993; Kister et al., 2003). Fluid flow modelling of larger convection cells has been the subject of a major collaborative research project (Ord, 2003) from which few results have been published.

As noted by Cuney et al. (2003) and Ramaekers et al. (2005a), Wolverine Point and Douglas formation mudstone should have acted as aquitards constraining such convection cells. Wilson et al. (2005) have shown that hydrocarbons sourced from the Douglas Formation may well have migrated throughout the basin, somehow crossing the Wolverine Point Formation in the process, however this for them was a post-ore process. Ramaekers et al. (2005a) have modelled the introduction of hydrocarbons as a result of late tilting of this shallow basin, thereby allowing the hydrocarbons to migrate "up-dip" into stratigraphically lower formations. The same "up-dip" process can be invoked for the introduction of Cretaceous hydrocarbons that permeate the Athabasca Group in Alberta. Kyser et al. (2000) reasoned that fluid flow events in the Athabasca Basin spanned more than 600 Ma, during much of that time under elevated temperatures approaching or at hydrothermal conditions (but to date, not higher than 240°C). It is not clear how many different fluid flow regimes might have been in place over that remarkable length of time.

Work by Jefferson et al. (2001), Hiatt and Kyser (2005) and the normal anisotropy of sedimentary layering indicate that the bulk of fluid flow within the basin was along sub-horizontal bedding planes and/or along the near vertical to 45° dipping fault zones. The convexity of leisegang patterns indicates fluid flow during early hematitization in the McArthur River area was toward the northeast, exiting the basin. Late limonitic banding documents recent meteoric fluid flow downward and outward from fault zones. Quirt (2003) reviewed evidence of fluid flow during mineralization that was either downward into basement faults (ingress) or upward and outward from basement fault zones (egress). Clearly any hydrothermal convection requires both of these processes to co-exist and be linked, and thus the downward, upward and lateral components of convection systems must have been overall balanced in terms of fluid volumes, heat flow etc. In any given ore-bearing fault system, fluids must have been entering the fault plumbing at one or more points at the same time as fluids exited elsewhere. Where the ingress or egress foci were prolonged, and suitable redox conditions existed, ore deposits were formed. In other places very large amounts of fluids were circulated and created large alteration zones including anomalous U contents in extensive halos, yet there was insufficient focus to form ore-grade concentrations. One of the challenges of regional versus detailed alteration analysis is to distinguish well focused from poorly focused fluid flow.

The triggers and drivers for hydrothermal mineralizing flow within intracontinental Mesoproterozoic basins has long been considered enigmatic (Ross, 2000) but clearly are a function of late-stage transpressive tectonic processes such as Ramaekers et al. (2005a) have begun modelling for the Athabasca Basin. Ruzicka (1996a) used terms such as "rapid subsidence" and "rifting" to describe events that triggered hydrologic systems, however such events were neither as "rapid" nor "rifting" as dramatic as in continental margin basins or foreland 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 continental 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-associated 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 formation. Compilations of fault arrays (e.g. Portella and Annesley, 2000a) are already complex, but this is only the beginning.

The authors agree with Cuney et al. (2003) regarding the uncertainty of ages and durations of alteration and ore deposition, and paraphrase their thoughts here. U-Pb ages on U oxides have errors of several to tens of Ma, attributed to diffusion of radiogenic Pb out of U-oxides (continuous or/and episodic) because of the large ionic radius of Pb. It is unknown whether the common ore ages of about 1350, 1000 and 300 Ma are all re-set from older primary ages of 1500 Ma or more that have recently been determined on the McArthur, Cigar Lake and Sue deposits (e.g. Alexandre et al., 2003; Cumming and Krstic, 1992; Fayek et al., 2002a, b). Calculation of the durations necessary to form these massive uraninite ore bodies requires numerous parameters, most of which are poorly constrained. Assuming that the mineralizing fluid contained 5 to 10 ppm U, percolated at rates in the order of 0.1 m/year, and had a volume of several tens of cubic km, the formation of Cigar Lake would have required a few million years. The Cigar Lake example is one where the ore forming process appears to have been relatively quiescent in that faults neither offset nor incorporate significant volumes of ore, nevertheless it is located directly above a basement shear zone, fracture zones are clearly the preferred sites of overlying perched and underlying basement-hosted ore lenses, and a basement ridge directly beneath the deposit has been interpreted as both a pre-Athabasca erosional remnant, and a post-Athabasca structural feature that was generated by local extensional tectonics (Andrade, 2002).

A very different approach can be taken to calculating the time of formation of basement-hosted deposits such as Sue C (as mapped by Tourigny et al. 2005) and McArthur River (as introduced by McGill et al. 1993). The styles of these two deposits are very similar, and distinct from that of Cigar Lake. Uraninite at these two deposits was deposited as lenses and pods within low-pressure dilatant jogs of transpressive fault regimes, and invokes very active processes such as seismic pumping (Tourigny et al., 2005; Sibson, 2001), also known as fault valve behaviour (Nguyen et al., 1998). These deposits probably formed in repeated increments during successive fault re-activation, and the duration of ore formation may be constrained by the duration of fault activity. In the case of the P2 Fault at McArthur River, the fault activity spanned from at least immediately before and during all of the deposition of the Read and Manitou Falls formations, i.e. the second sequence of the Athabasca Group (Ramaekers et al. 2005b). The end of this activity in the McArthur River mine area cannot be determined because the present day erosional surface truncates both the Manitou Falls Formation and the alteration halo.

Typical second-order basin-filling sequences are thought to require 22-45 million years for deposition (Krapez, 1996); nevertheless, the duration of Precambrian examples is difficult to determine with precision. Basement and detrital zircon geochronology constrain the start of Athabasca Group sedimentation in the McArthur River area to 1740-1730 Ma (Rainbird et al., 2005). An age of 1644±13Ma (U-Pb, Rainbird et al. 2005) on intraformational volcanic ash zircons in Wolverine Point Formation (Sequence 3) is close to U-Pb dates on fluorapatite in Sequences 1 and 3 (Cumming et al., 1987). These dates allow approximately 100 Ma for deposition of Sequences 1 and 2. Deposition of upper Sequence 3 and lower Sequence 4 could have taken place at any time during the next 100 Ma, as constrained by Re-Os dating by Creaser et al. (2005, in prep.) on organic-rich shale of the Douglas Formation (second highest preserved unit of the Athabasca Group).

Given the above stratigraphic context, the age of ore-related alteration and the oldest U deposit is either synchronous with or post-dates deposition of Sequence 3, and is similar in age to the widespread diagenetic fluorapatite in the Athabasca Basin. The U/Pb and 40Ar/39Ar dating of uraninite and clays indicate that pre-ore alteration occurred between 1620 Ma and 1670 Ma, followed by initial U mineralization between 1550 and 1600 Ma (Alexandre et al. 2003), predating deposition of the Wolverine Point Formation. More conservative U-Pb dating has provided oldest ages of 1486-1519 Ma on the McArthur River and Sue C deposits (Fayek et al., 2002b) and 1461 ± 47, 1176 ± 9, and 876 ± 14 Ma (± 1r) Ma on the Cigar Lake deposit (Fayek et al., 2002a) - these post-date the Wolverine Point Formation and pre-date the Douglas Formation. Older uraninite ages of about 1640 Ma have been determined in the Kombolgie Basin (Ludwig et al. 1987; Maas 1989) and suggested by paragenesis in the Thelon Basin (Kyser et al. 2000). Clearly much more work remains to better constrain the durations and timings of these deposits, as advocated by Cuney et al. (2003) and Kyser et al. (2000).

Quartz dissolution, locally up to 90 %, is spatially associated with U-mineralization at deposits such as Cigar Lake (Andrade, 2002) as well as with breccia bodies (Lorilleux, 2000), and is considered another process that created space for ore (Cramer, 1986; Cuney et al. 2003). It has also been suggested that massive dissolution of rock in such alteration zones could have contributed U to form deposits such as Dawn Lake and Cigar Lake (Cramer, 1986; Percival, 1989; Kyser et al., 2000). The silica-undersaturated fluid required to dissolve the quartz could not have been the diagenetic basin brine because this was in equilibrium with ubiquitous framework quartz grains constituting the Athabasca Group. Cuney et al. (2003) considered that the upward decrease of alteration intensity from the unconformity and the input of basement-derived elements (Mg for sudoïte and dravite, K for illite) indicate that these fluids were derived from the basement. The mechanism responsible for silica under-saturation has not been identified (work in progress, M. Cuney, pers. comm. 2005) but this phenomenon can be compared to the mineralization in episyenite columns that developed in large bodies of quartz-rich granitoid rocks (Cuney et al., 2003), some of which contain U deposits like the Gunnar Mine of the Beaverlodge District (Gandhi, 1983). Dissolution of quartz from the granitoid rocks created space in which albite was precipitated (along with uraninite) to convert the rock to syenite (hence, episyenite).

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 sandstones, and Ni, Co, Cu, Zn, Au in the polymetallic deposits. However, evidence for such a reduced fluid is still very weak. Experimentally, pure graphite and pure water do not react below 400°C, and Wilson et al. (2005) consider such a reaction to be thermodynamically impossible in any quantitative sense. Cuney et al. (2003) nevertheless propose further experimental studies to test the reactivity of disorganized graphitic grains from the basement with highly chlorinated Na-Ca brines at low temperature. Ongoing studies by Cuney et al. (2003) in the Athabasca and Kombolgie U districts have identified CH_4 and N_2 in the gas phase of some highly saline inclusions and in gas-rich inclusions in the basement gneiss near ore deposits. Hydrocarbons reported in the Rabbit Lake deposit (Pagel and Jaffrezic, 1977) and Nabarlek and Jabiluka deposits (Wilde et al., 1989) have only been detected by quartz crushing and global analyses. These observations are the only indications of a reduced fluid. We suggest here that pursuing graphite in the geochemical reactivity sense may be of limited value (but we have to prove if it is able to react or not), and consideration of sulfide minerals and electrochemical processes (good for gold deposition but how to reduce U(VI) to U(IV) with this process?) may prove to be more fruitful lines of research.

Although there is litttle doubt that graphite has been removed in the altered portions of graphitic shear zones that underlie uranium deposits (e.g. Landais et al., 1993), how that graphite was removed, and the origin of the hydrocarbons and of the bitumen found with and as globular coatings around uraninite continue to be debated. For example, Sangely et al. (2003) noted the similar isotopic composition of barren bitumens and basement graphite, McCready et al. (1999) postulated bitumens as a key precursor to uraninite ore, and Annesley et al. (2001) described intimate textural intergrowths between graphite and hydrocarbons as analysed by synchrotron methods. Nevertheless, straightforward paragenetic studies on deposits spanning the Athabasca Basin clearly show that virtually all hydrocarbon material texturally post-dates uraninite and structurally comminuted graphite (Wilson et al. 2005). Any minor amounts of uraninite that are claimed to envelop bitumen can be discounted as probably remobilised (unless they can be precisely dated). Because pyrobitumen is either absent or sparse in a number of high-grade deposits (e.g. McArthur River), it is clearly not a necessary condition for the presence of ore. Bitumen still has some exploration significance as a post-ore accessory that is commonly present, however in the western part of Athabasca Basin, such bitumen is a pervasive pore-filler that results from massive invasion of the Athabasca Basin by the same material that now constitutes the Cretaceous Athabasca oil sands. We view the hydrocarbon as having been precipitated by contact with the highly reactive uraninite, and congealed partly through radiation damage to the organic molecules. (See Sangély et al., 2003 for a similar process but with an abiogenic origin of the organic matter).

New Areas Of High Uranium Potential In Canada

The Athabasca and Thelon basins remain the areas in Canada with the most potential for new discoveries of unconformity-associated U deposits. Whereas it might seem that exploration is at a mature stage, particularly in the Athabasca Basin, this single basin is larger than some provinces and many countries of the world, and only a small part of it has been touched by intensive exploration. The entire basal unconformity surface is prospective where it intersects favourable basement domains (such as the Wollaston-Mudjatik transition domain of the eastern Athabasca Basin) and re-activated graphitic shear zones, but it is all hidden except around the well-explored rim (Figures 2 and 3). The entire Athabasca and Thelon basins would benefit from high-resolution airborne geophysical surveys as follows: a) magnetics for better interpretations of basement geology, b) multiparameter gamma ray to map in detail the distribution of U, Th and K that are fundamental measures of regional to detailed alteration zones, c) electromagnetics to capture all of the potential graphitic conductors, and d) gravity to delineate minor fault offsets of the basal unconformity beneath sedimentary cover. Integrated multiparameter geophysical and geological transects like that completed by EXTECH IV across the McArthur River deposit camp (Jefferson et al. 2002b, 2003a, b, c) might provide the common road maps required to assess undiscovered resource potential in these vast under-explored areas.

Blind, ingress-type, basement-hosted, monomineralic deposits are of particular interest in the Athabasca Basin. These are extremely difficult to find with our current technology, but are attractive targets because many likely remain to be found in areas of shallow Athabasca Group cover, or even outside the basin. Nisto, a past producer northeast of Black Lake (Macdonald et al., 2000) is one possible example. Such deposits are suitable for open pit mining, avoiding the need for freeze wall technology to control water, and they are metallurgically attractive.

If both ingress and egress type deposits are located along reactivated basement fault systems, and if hydrothermal convection is integral to the genesis of unconformity-associated deposits, each fault system that hosts an egress type deposit has potential to host ingress-type deposits. Geophysical, geochemical and mineralogical tools need to be developed to locate sites of focused downwelling where ingress deposits might be found. Re-evaluation of historical exploration, particularly drilling, must be 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.

Renewed study of the Great Bear vein U district and its relationship to the Hornby Bay Group may provide insight as to their mutual age and spatial relationships, and test the notion that these vein deposits are exhumed unconformity-associated deposits that were originally beneath an extension of the Hornby Basin. Such new information might provide foci for renewed exploration of the Hornby Basin.

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