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4 The Millennium uranium deposit, Athabasca Basin, Saskatchewan, Canada: an atypical
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6 basement-hosted unconformity-related uranium deposit.
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Abstract

The Millennium uranium basement-hosted unconformity-related deposit is one of the most recent discoveries of basement-hosted deposits, and may hold as much as 46.8 million pounds (18 000 metric tonnes U) at an average grade of 4.53% U_3O_8 . The alteration zone associated with the deposit is atypical, as it consists predominantly of muscovite rather than the chlorite and muscovite commonly found in the other basement-hosted deposits of the Athabasca Basin.

The pre-ore stage consists of early minor clinochlore alteration at temperatures around 275°C interpreted to result from retrograde metamorphism of the basement rocks prior to the deposition of the Athabasca Basin. This clinochlore event is followed by an extensive 1Mc muscovite alteration of the basement rocks by Na-K-Fe-rich basinal brines at temperatures around 250°C, creating up to 25% void space in the basement rocks. Subsequently, a weak chamoisite alteration resulted from Fe-Mg-Ca-rich fluids with temperatures around 300°C that evolved from reactions between the basinal brines and the basement rocks. Ore stage uraninite has a U-Pb date of 1590 Ma and is contemporaneous with syn-ore 1Mc muscovite and iron oxide deposited at temperatures near 250°C from fluids with similar $\delta^{18}\text{O}$ but lower δD values to pre-ore basinal fluids. The low δD values indicate post-depositional exchange of H isotopes with Cenozoic fluids. Post-ore alteration consists of dravite and sudoite precipitated from Mg-Ca rich fluids, from the same evolved basinal brines that precipitated pre-ore chamoisite, at temperatures around 185°C, filling voids created during the pre-ore stage. Our results indicate that the pre- and syn-ore minerals were formed dominantly by basinal fluids, with no contribution from basement fluids, which differ from some other basement-

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4 hosted deposits in the Athabasca Basin where basement-derived fluid influx was more
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6 significant.
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9 Several U-Pb and Ar-Ar resetting events recorded on pre-ore muscovite and on
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11 uraninite are coincident with fluid flow induced by distal orogenies, remobilizing
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13 radiogenic Pb from the deposit. The radiogenic Pb remobilization events were not
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15 widespread as the presence of radiogenic Pb is restricted to basement rocks proximal to
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17 the deposit and along the unconformity.
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23 **Introduction**

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25 Paleoproterozoic sandstone- and basement-hosted unconformity-related uranium
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27 deposits in the Athabasca Basin, Canada, account for about one third of the global stock
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29 of uranium (McGill et al., 1993; Krasenberg, 2004; Kyser and Cuney, 2008). There are
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31 two general end-member types of unconformity-related deposits: those hosted primarily
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33 in the sandstone above the unconformity (sandstone-hosted or egress-style) and those that
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35 occur in the metamorphic basement rocks below the unconformity (basement-hosted or
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37 ingress-style) (e.g. Jefferson et al., 2007; Kyser and Cuney, 2008). In general, alteration
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39 associated with basement-hosted deposits consists of an outer illitic halo, an inner illitic-
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41 chloritic halo and a core containing most of the uranium mineralization (e.g. Fayek and
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43 Kyser, 1997; Thomas, 1998; Alexandre, 2005). Zones of silicification and dravitzation
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45 are commonly observed above the mineralized zones (Thomas et al., 1998) and
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47 hydrothermal assemblages in sandstone units overlying basement-hosted deposits include
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49 quartz, kaolinite, illite, chlorite and dravite (McGill, 1993). Basement-related
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51 mineralization in most cases is in close proximity to graphitic units related with pre-
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Athabasca Basin steeply to moderately dipping fractures and breccias, which may extend down to 400 meters into the basement (Thomas et al., 1998). Mineralization typically occurs as disseminated and massive uraninite, with an average grade of 1 to 3% U_3O_8 , although McArthur River has a much higher average grade of 14% U_3O_8 (Thomas et al., 1998).

Two distinct models have been proposed to explain the source of uranium in basement-hosted deposits. One model suggests that the source of uranium is mainly from uranium-rich phases, such as apatite, zircon and monazite from the sandstones of the Manitou Falls Formation. In this model, uranium is leached from these minerals by basinal oxidizing fluids and transported downwards into suitable structural traps such as a fracture zones and faults in the basement, and then deposited by reaction with reducing lithologies or by mixing with reducing fluids of basement origin (e.g. Wilson and Kyser, 1987; Kotzer and Kyser, 1995; Fayek and Kyser, 1997). However, Hecht and Cuney (2000) and Derome et al. (2003) propose that the source of uranium is monazite from the metamorphic basement rocks and that uranium is leached from these phases by basin-derived oxidizing brines, and then precipitated below or at the unconformity by mixing with reduced fluids of unknown origin or by interacting with Fe-bearing rocks.

The Millennium deposit, which was discovered in 2000 by Cameco Corporation (Roy et al., 2005; Fig. 1), occurs on a major post-Athabasca Group sandstone structural trend in the eastern Athabasca Basin, in close proximity to a graphitic horizon at a depth of 650 meters from the surface and 100 meters below the unconformity (Roy et al., 2005; Figs. 2a, b, 3a). Present indicated resources are 46.8 million pounds (18 000 metric tonnes U) at an average grade of 4.53% U_3O_8 (Thomas, 2002; Roy et al., 2005). The

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4 mineralization at Millennium is situated in the footwall of a major reverse fault and is
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6 broadly stratabound, although several styles of mineralization are observed (Fig. 3a). The
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8 mineralization is essentially monometallic, with Pb and V contents increasing with
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10 uranium enrichment (Roy et al., 2005). Manitou Falls Formation sandstones above
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12 Millennium have the typical alteration observed in the sandstone above other basement-
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14 hosted deposits, with an envelope of illite, chlorite, kaolinite, dravite and dickite (McGill,
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16 1993; Roy et al., 2005). The most striking feature at Millennium is the mineralogy of the
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18 alteration envelope that surrounds the deposit, as it comprises mainly of illite, which is
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20 atypical of most basement-hosted deposits in which chlorite and illite are the dominant
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22 alteration minerals (Fig. 3b). Minor amounts of chlorite, dravite, kaolinite and fine
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24 grained hydrothermal quartz occur throughout the alteration zone of the Millennium
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26 deposit.
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33 In this paper, we report detailed petrographic, mineralogical and isotopic data
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35 from the Millennium deposit to constrain the mechanisms responsible for the formation
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37 of this atypical alteration zone and the uranium mineralization. Based on these results, we
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39 propose a model for the genesis of the deposit, with implications for exploration of
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41 similar deposits.
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50 **Geology**

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52 The Millennium deposit is situated in the Paleoproterozoic Athabasca Basin, at 45
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54 km from its eastern margin, 35 km southwest of the McArthur River deposit and 35 km
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56 north of the Key Lake deposit (Fig. 1). The deposit is hosted in basement rocks in the
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4 transition zone between the Archean and Paleoproterozoic rocks of the Wollaston
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6 Domain and the Mudjatik Domain, both belonging to the Cree Lake Mobile Zone of the
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8 Hearne Province (Hoffman, 1988; Tran et al., 2003; Fig. 1).
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11 The Wollaston Domain is a northeast-trending fold and thrust belt that is fault-
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13 bounded to the east by the Peter Lake Domain and to the west by the Mudjatik Domain
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15 (Money, 1968; Fig. 1). Three main groups of rocks are distinguished (Lewry and Sibbald,
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17 1977): (1) 2800-2950 Ma Archean granitic, granodioritic, and tonalitic orthogneisses and
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19 subordinate metamorphic rocks (Annesley et al., 2005), (2) unconformably overlying
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21 ~1920 Ma high-grade Paleoproterozoic graphitic and non-graphitic pelitic,
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23 psammopelitic, and psammitic gneisses, subordinate metaquartzite, calc-silicates, and
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25 amphibolites, as well as rare BIF of the Wollaston Group (Lewry and Sibbald, 1980; Tran
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27 et al., 2003), (3) 1810-1840 Ma deformed granitoids, minor gabbros and pegmatites
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29 (Annesley et al., 2005).
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36 The Mudjatik Domain is a northeast-trending fold and thrust belt that is fault-
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38 bounded to the east by the Wollaston Domain and to the west by the Virgin River
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40 Domain (Money, 1968; Fig. 1). It comprises rocks similar in composition to the
41
42 Wollaston Domain, however, it is dominated by granitoid gneisses derived in part from in
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44 situ migmatization and anatexis (Lewry and Sibbald 1980) and contains a larger volume
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46 of late peraluminous granite (Roy et al., 2005).
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51 The Wollaston-Mudjatik transition zone is characterized by a structurally
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53 complex zone comprising elongated Archean granitoid domes (mega-boudins), thrust-
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55 and strike-slip structures and related duplex structures (Annesley et al., 2005). The
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57 boundary is defined on the basis of lithological changes in the Archean basement,
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4 lithofacies changes in the Wollaston Group, and changes in types of Hudsonian intrusive
5 rocks. The boundary is a major anastomosing northeast-trending transcurrent shear zone
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7 associated with a thermotectonic front, and is defined as a transitional zone from thick-
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9 skin tectonics in the Wollaston Domain to thin-skin tectonics in the Mudjatik Domain
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14 (Annesley et al., 2005).

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16 The rocks of the Wollaston and Mudjatik Domain were complexly deformed and
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18 metamorphosed to amphibolite facies by the Trans-Hudson Orogen, which reached peak
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20 metamorphism at ca. 1800–1820 Ma (Lewry and Sibbald, 1980; Kyser et al., 2000).
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22 Subsequent rapid uplift initiated the formation of the Athabasca Basin at maximum age
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24 of 1750 Ma, as recorded by K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages (Burwash et al., 1962;
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26 Kyser et al., 2000; Alexandre et al., 2009), and at a minimum age of 1620–1640 Ma
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28 obtained on early diagenetic fluoroapatite (Rainbird et al., 2003). The Athabasca Basin
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30 formed as a series of three NE–SW oriented sub-basins, with the easternmost Cree sub-
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32 basin hosting the majority of the uranium deposits (Armstrong and Ramaekers, 1985;
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34 Kyser et al., 2000) and sedimentation continued until the deposition of the carbonaceous
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36 marine shale of the Douglas Formation at ca. 1540 Ma (Creaser and Statiuk, 2007).
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43 Sediments of the Athabasca Basin consist of sequences of Paleoproterozoic and
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45 Mesoproterozoic polycyclic, mature fluvial to marine quartz sandstone, collectively
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47 referred to as the Athabasca Group. These were deposited in a near-shore shallow shelf
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49 environment (Ramaekers and Dunn, 1977; Ramaekers, 1990). The Athabasca Group in
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51 the Eastern part of the Athabasca Basin comprises exclusively the Manitou Falls
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53 Formation, which consists of coarse to fine-grained hematite-rich conglomerates along
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55 thin stratigraphic horizons, indicating oxidation of heavy mineral layers and silty
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4 sandstones filling (Ramaekers, 1990). Ramaekers et al. (2007) proposed a revised
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6 stratigraphy of the Athabasca Group introducing newly recognized Smart Formations in
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8 the western Athabasca and Read Formation in the eastern Athabasca, which replace the A
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10 Member of the Manitou Falls Formation. Ramaekers et al. (2007) also subdivide the
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12 Manitou Falls Formation, introducing the Warnes Member (MFw), Raibl Member (MFr)
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14 and Bird Member (MFb), that are the equivalent of the Manitou Falls B Member, but are
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16 attributed to the Karras deposystem for the Warnes Member, Moosonee deposystem for
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18 the Raibl Member and the Ahenakew deposystem for the Bird Member. To be consistent
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20 with past publications and industry practices, this paper follows the original stratigraphy
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22 proposed by Ramaekers (1990) and will refer to the Read Formation as the Manitou Falls
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24 A Member and the Warnes, Raibl and Bird members of the Manitou Falls Formation as
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26 the Manitou Falls B Member.
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33 The Manitou Falls Formation and underlying basement are cut by a series of
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35 mafic dikes known as the McKenzie dike swarms (Cumming and Kristic, 1992), dated at
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37 1267±2 Ma from U–Pb date from baddeleyite (LeCheminant and Heaman 1989). They
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39 range from 1 m to several hundred meters wide, and are controlled by tensional trends
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41 associated with left-lateral movement along the ancient Hudsonian faults (Hoeve and
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43 Sibbald, 1978; Sibbald and Quirt, 1987).
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48 The Millennium deposit is located within Paleoproterozoic metasedimentary
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50 basement rocks of the Wollaston-Mudjatik transition zone at a depth ranging from 600 to
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52 more than 750 meters from the surface, and from 100 to more than 150 meters below the
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54 sandstone-basement unconformity (Figs. 1, 2a, b, c, 3a). The deposit is interpreted to lie
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56 along the northwest-trending limb of a Z-type parasitic fold structure associated with the
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4 east limb of a larger regional north-plunging synformal fold that closes to the south-
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6 southwest (Thomas, 2002). The mineralization is located between two reverse faults and
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8 is essentially stratabound (Fig. 3a). The footwall reverse fault, known as the Mother fault,
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10 is filled with pre-ore quartz, and the hanging wall reverse fault is associated with a
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12 cordierite-graphite metapelitic gneiss unit closely associated with the mineralization (Fig.
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14 3a). Basement rocks consist of graphitic metapelitic schists and gneisses and lesser
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16 metacalcopelitic rocks that have been intruded by numerous pegmatitic and granitic rocks
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18 (Fig. 3a; Thomas, 2002). They are overlain by the Paleoproterozoic Athabasca Group
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20 which consists of 500-700 meters of diagenetically altered A, B, C, and D members of
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22 the Manitou Falls Formation (Ramaekers 1990; Roy 2005) cut by fractures and faults.
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24 The mineralization is hosted within the metapelitic gneisses and is overprinted by
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26 extensive illite alteration (Roy et al., 2005; Fig. 3b). The alteration pattern was
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28 interpreted by Thomas (2002) to reflect a zonation from distal saussuritization
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30 (plagioclase retrogression) and sericitization, through a more proximal zone of
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32 chloritization into a central zone of increasing illitic and dravite alteration. In contrast,
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34 other major basement-hosted deposits are characterized by a small outer illitic halo, a
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36 large inner chloritic halo and a core containing most of the uranium mineralization (e.g.
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38 Fayek and Kyser, 1997; Thomas, 1998; Alexandre, 2005).
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48 Uranium mineralization consists predominantly of uraninite with minor amounts
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50 of coffinite (Roy et al., 2005) and occurs in a variety of styles including massive
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52 foliation-controlled replacement, uraninite matrix in breccias, irregular fracture-
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54 controlled infillings and thin vein-type uraninite, bleb-like aggregates and thin discordant
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56 uraninite veinlets and rims around fragments or quartz veins (Thomas, 2002). Massive
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4 replacement type is the dominant style of mineralization, while fracture infilling and
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6 vein-type is less well developed (Thomas, 2002). Textures in breccia-hosted
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8 mineralization suggest a progression from simple fracture-controlled, through more
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10 complex hydraulic fracturing, and finally into corrosive solution breccias characterized
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12 by progressive rounding and embayment of clasts (Thomas, 2002). Width and grade of
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14 the ore zones are variable with grades ranging between 1 and 4% U over 20 to 30 meters
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18 (Roy et al., 2005).
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23 **Methodology**

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26 A total of 84 samples were taken from 15 drills holes delineating a N-S transect in
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28 the Manitou Falls Formation sandstone and a E-W transect in the Manitou Falls
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30 Formation sandstone and underlying basement (Figs. 2a, b, 3a). Polished thin sections of
31
32 all samples were examined using conventional transmitted and reflected light microscopy
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34 to determine mineral paragenesis. All samples were analyzed with a Portable Infrared
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36 Mineral Analyzer (PIMA) in order to conduct preliminary identification of the hydrous
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38 minerals. The samples were crushed and sieved and clay minerals were extracted from
39
40 the coarsest crushed fraction (>1.4 mm) by ultrasound disintegration. X-ray diffraction
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42 (XRD) analyses were performed on all size fractions of the mineral separates obtained
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44 from ultrasound disintegration (<2 μm , 2-5 μm and >5 μm) using a Seimens X-Pert at
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46 Queen's University, Kingston, Canada. The <2 μm pure monomineralic and mixed clays
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48 separates were selected for stable isotope analyses and 2-5 μm pure monomineralic
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50 separates of muscovite were analyzed by $^{40}\text{Ar}/^{39}\text{Ar}$ techniques. Chemical compositions of
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52 alteration minerals were determined by electron microprobe analyses (EMPA) using a
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4 Camebax MBX electron microprobe equipped with 4 wavelength dispersive X-ray
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6 spectrometer (WDX) at Carleton University, Ottawa, Canada. Analyses are accurate to 1-
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8 2 % relative for major elements (>10 wt %), 3-5 % relative for minor elements (>0.5 -
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10 <5.0 wt %). As detection limit is approached (<0.1 wt %), relative errors approach 100
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12 %.

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14 The compositions of chlorite and muscovite were used to estimate their temperatures
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16 of formation. The chemical composition of chlorite and muscovite was used to estimate
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18 temperatures of formation based on tetrahedral site occupancy (Cathelineau, 1988) and
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20 are accurate to within 30°C.

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23 For stable isotopes, oxygen was extracted using the BrF₅ method of Clayton and
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25 Mayeda (1963) and $\delta^{18}\text{O}$ values were measured using a dual inlet Finnigan MAT 252
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27 isotopic ratio mass spectrometer. Hydrogen isotopic compositions were determined using
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29 a TC/EA Thermo Finnigan and a DeltaPlus XP Finnigan Mat mass spectrometer. O and
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31 H isotopic compositions are reported in the δ notation in units of per mil relative to the
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33 standard V-SMOW. The $\delta^{18}\text{O}$ and δD values were reproducible to ± 0.2 and ± 3 permil,
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35 respectively. Oxygen isotope fractionation factors used throughout this paper are those
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37 proposed by Wenner and Taylor (1971) for water-chlorite, O'Neil and Taylor (1969) for
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39 water-muscovite and Fayek and Kyser (2000) for water-uraninite. Hydrogen isotope
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41 fractionation factors used are those proposed by Marumo et al. (1980) for water-chlorite
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43 and Vennemann and O'Neil (1996) for water-muscovite. In previous studies, the water-
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45 illite fractionation factor was used instead of water-muscovite. Hydrogen water-illite
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47 fractionation factors of Yeh (1980) used in previous studies were calibrated in the Texas
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49 Gulf Coast and used a natural diagenesis progression from kaolinite to smectite to
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51 illite/smectite to illite. The illite chemical composition from the Texas Gulf Coast varies
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4 between 2-6% K₂O and 2- 6% FeO (Awwiller, 1993) and is different to the chemical
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6 composition of the white mica present in the Athabasca, where K₂O generally varies
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8 between 8 and 10% and FeO is usually <1% (Wilson and Kyser, 1987; Kotzer and Kyser,
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10 1995; Alexandre et al. 2005). Moreover, Vennemann and O'Neil (1996) pointed out that
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12 Fe content has the strongest effect on the fractionation of hydrogen-mineral, wherein the
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14 water-mineral fractionation factor increases with decreasing Fe content. The δD values of
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16 the fluids are 38‰ higher at 250°C using the muscovite-water fractionation factors
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18 relative to those using illite-water. Therefore we prefer the muscovite-water fractionation
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20 factor of Vennemann and O'Neil (1996) as the Fe content is closer to what is observed in
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22 muscovite from the Athabasca Basin.
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28 U-Pb isotope ratios were determined using LA-HR-ICP-MS (Chipley et al., 2007)
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30 using a Finnigan MAT ELEMENT HR-ICP-MS and a NEPTUNE HR-MC-ICP-MS,
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32 both equipped with a high-performance Nd:YAG New Wave UP-213 laser ablation
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34 system at the Queen's Facility for Isotope Research (QFIR). Ablation of uraninites was
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36 achieved on polished thin sections using a 35-40 μm spot size with 35-40% laser power
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38 at a frequency of 2Hz. The gas (argon in all cases) flows were as follows: cooling gas, 15
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40 l/min; auxiliary gas, 1.0 l/min; sample carrier gas, 1.0 l/min. A low mass resolution of
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42 350 (defined as the ratio of mass over peak width at 5% of the signal height) was used.
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44 For each sample, ²⁰⁴Pb, ²⁰⁶Pb, ²⁰⁷Pb, ²³⁵U and ²³⁸U were measured and corrections for
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46 common Pb were made scan-by-scan to each spot. No corrections were made to the
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48 ²³⁸U/²³⁵U ratios as they were near the 137.88 natural ratio.
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55 ⁴⁰Ar/³⁹Ar dating was performed on two pure monomineralic separates of
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57 muscovite using the 2 to 5 μm size fraction. Plateau ages were calculated using not less of
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70% of the gas released and three consecutive steps that overlap in their 1σ error margin while pseudo-plateau ages are defined by 30-70 % of the gas released. The $^{40}\text{Ar}/^{39}\text{Ar}$ dating was performed at the Noble Gas Laboratory, Pacific Centre for Isotopic and Geochemical Research, University of British Columbia. Mineral separates were wrapped in aluminum foil and stacked in an irradiation capsule with similar-aged samples and neutron flux monitors (Fish Canyon Tuff sanidine (FCs), 28.02 Ma (Renne et al., 1998). The samples were irradiated at the McMaster Nuclear Reactor in Hamilton, Ontario, for 90 MWH, with a neutron flux of approximately 6×10^{13} neutrons/cm²/s. Analyses (n=45) of 15 neutron flux monitor positions produced errors of <0.5% in the J value. The mineral separates were step-heated at incrementally higher powers in the defocused beam of a 10W CO₂ laser (New Wave Research MIR10) until fused. The gas evolved from each step was analyzed by a VG5400 mass spectrometer equipped with an ion-counting electron multiplier. All measurements were corrected for total system blank, mass spectrometer sensitivity, mass discrimination, radioactive decay during and subsequent to irradiation, as well as interfering Ar from atmospheric contamination and the irradiation of Ca, Cl and K (Isotope production ratios: $(^{40}\text{Ar}/^{39}\text{Ar})_{\text{K}}=0.0302 \pm 0.00006$, $(^{37}\text{Ar}/^{39}\text{Ar})_{\text{Ca}}=1416.4 \pm 0.5$, $(^{36}\text{Ar}/^{39}\text{Ar})_{\text{Ca}}=0.3952 \pm 0.0004$, $\text{Ca/K}=1.83 \pm 0.01$ ($^{37}\text{Ar}_{\text{Ca}}/^{39}\text{Ar}_{\text{K}}$).

Forty-one Manitou Falls Formation sandstones and fourteen basement samples were analyzed for leachable Pb and U isotopes and leachable trace elements. Samples were crushed and then leached following the technique of Holk et al. (2003). The 0.50–1.40 mm crushed fraction was used and about 0.5 g of sample and 5 ml of 0.02 M HNO₃ were loaded into a polyurethane tube, placed in an ultrasonic bath for 120 min and centrifuged. One gram of the liquid was diluted with 100 g of 1% HNO₃ spiked with ^{115}In

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4 and the Pb isotopic ratios and trace element concentrations were measured using a
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6 Finnigan MAT ELEMENT high-resolution ICP-MS. The Pb and U isotope ratios were
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8 calculated using the signal intensities (counts/s) in low-resolution mode and have an
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10 average uncertainty of ca. 1% based on repeat analyses. Corrections were made for
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12 interferences from Hg and mass fractionation was monitored using Tl in the solutions and
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14 externally with in-house and NIST Pb isotope standards (NBS 981, normal Pb and NBS
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16 983, radiogenic Pb). For each sample, ^{201}Hg , ^{202}Hg , ^{204}Pb , ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{235}U and
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18 ^{238}U were measured, blank subtracted and ^{204}Pb was corrected for ^{204}Hg interference
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20 using ^{202}Hg . Concentrations were corrected for instrument drift and matrix using ^{115}In as
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22 an internal standard and external calibration for element concentrations and blank
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24 subtraction. $^{206}\text{Pb}/^{204}\text{Pb}$ ratios greater than 30 were considered radiogenic.
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35 **Paragenesis of Alteration Minerals**

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37 Mineral paragenesis for the sub-Athabasca basement and the Manitou Falls
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39 Formation sandstones based on textural relationships are shown in Figures 4a and 4b,
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41 respectively. Three main stages of alteration were recognized in the basement (pre-ore,
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43 syn-ore and post-ore) and four main stages of alteration were recognized in the basin
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45 (diagenesis, early and late hydrothermal alteration and late alteration).
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52 *Basement alteration*

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54 Detailed basement host rock descriptions by Roy et al. (2005) described ten
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56 stratigraphic units consisting of metapelitic to metasemipelitic to metacalcipelitic gneisses
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58 associated with minor graphite, a cordierite-graphite metapelitic gneiss unit, granites,
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pegmatites and minor amphibolites (Figs. 3a, 4a). Mineral assemblages in the metapelitic to metasemipelitic gneisses consist of fine-grained quartz, plagioclase, biotite, rutile and minor amount of graphite (Fig. 5a). Metacalcipelitic gneisses contain the same assemblage as metapelitic to metasemipelitic gneisses with the addition of carbonate minerals. The cordierite graphitic metapelitic gneiss consists of fine-grained cordierite, graphite, quartz, plagioclase, biotite and rutile. The granite and pegmatite units consist of medium to coarse-grained quartz, plagioclase, potassium feldspar, muscovite and biotite. One small layer of amphibolite (~1.5 m) was identified and consists of hornblende and plagioclase.

Early pre-ore alteration of the metamorphosed basement metapelitic assemblages is characterized by C1 chlorite alteration of fine to medium-grained (0.5-10 mm) biotite within the metapelites (Figs. 4a, 5b) and recrystallization of rutile needles present within the biotite crystal to rutile aggregates in the chlorite crystal (Fig. 5b). This relationship is only observed in the distal alteration zone, which is >75 meters from the ore zone. Subsequently, these assemblages were overprinted by pervasive M1 muscovite alteration creating medium- to fine-grained anhedral muscovites (Fig. 4a, 5c). This muscovite alteration is observed proximal to the mineralization, but extends into the distal alteration zone up to 100 meters from the ore zone. Prior to syn-ore alteration is minor replacement of M1 muscovite to dark green C2 chlorite in the footwall of the fault associated with the cordierite-graphite metapelitic gneiss unit (Fig. 5d).

Syn-ore alteration of the basement rocks consists of zones of massive uraninite 5 to 10 meters below the cordierite-graphite metapelitic gneiss unit of the hanging wall fault (Figs. 3a,b, 4a, 5e) and zones of disseminated uraninite in the hanging wall of the

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4 Mother Fault (Fig. 5f). Uraninite has a finely mottled and speckled appearance, forms
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6 euhedral (Fig. 5e), colloform or anhedral grains (Fig. 5f), and is typically zoned in
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8 reflected light and back scattered electron images. Lower reflectivity or porous zones,
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10 particularly those in close proximity to the rims of the uraninites, are interpreted to
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12 indicate recrystallization during post-ore alteration stages. Uraninite is spatially
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14 associated or rimmed with medium- to coarse-grained (10-300 μm), anhedral to euhedral
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16 M2 muscovite and fine-grained H1 hematite (Fig. 5e, f). Some rare fine-grained euhedral
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18 aluminium phosphate-sulfate (APS) minerals occur with the disseminated mineralization
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20 within M2 muscovite aggregates.
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26 Post-ore alteration is characterized by the occurrence of minor dravite as fine-
27
28 grained acicular dravite aggregates (50-300 μm) and as veinlets of dravite $\pm$ quartz (50-
29
30 500 μm) overprinting the pre- and syn-ore assemblages throughout the alteration zone
31
32 (Fig. 5f). Locally, brecciated zones 25 to 40 meters wide containing a matrix of fine-
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34 grained dravite occur in the hanging wall of the Mother Fault (Fig. 3b). Small amounts of
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36 acicular coffinite (Fig. 6a) rim uraninite crystals in the ore zone and are interpreted to
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38 have formed between the dravite and later calcite precipitation. Minor amounts of thin
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40 (50-300 μm) calcite $\pm$ Py1 pyrite veins cut the dravite aggregates and veinlets and H2
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42 hematite partially replaces H1 hematite. Fine-grained C3 chlorite (5-100 μm ; Fig. 5f)
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44 with minor Py2 pyrite occurs in corroded zones filling the voids created by prior
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46 alteration, and is one of the late phases. Locally, C4 chlorite has altered medium-grained
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48 M1 muscovite in the fault associated with the cordierite-graphite metapelitic gneiss unit
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50 and is interpreted to have been formed during the late post-ore stage based on
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52 geothermometric results.
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Alteration of Manitou Falls Formation Sandstones

The earliest diagenetic textures preserved in the sandstones are thin veneers of red-brown SH1 hematite and poor to well developed syn-compaction SQ1 quartz overgrowths on detrital SQ0 quartz grains followed by fine-grained SK1 dickite (Fig. 4b, 7a, b, c). Fine-grained SM muscovite (Fig. 7a, b, c) occurs filling corroded zones in quartz overgrowths, characteristic of early hydrothermal alteration minerals throughout the Manitou Falls Formation sandstones and indicative of significant water/rock ratios (Wilson and Kyser, 1987; Kotzer and Kyser, 1995). Minor amounts of fine-grained euhedral APS minerals (SAPS) are observed to have crystallized simultaneously with early hydrothermal alteration SM muscovite. SK1 dickite and SM muscovite are partially replaced by needle-shaped ST dravite near the unconformity (Fig. 7c), which is interpreted to have formed during late hydrothermal alteration. Dravite occurs in minor amounts in all areas, except near the unconformity in drill hole CX56 where it is the only alteration mineral present in the sandstones interstices. Locally, SH2 hematite occurs in embayments in the margins of late hydrothermal alteration tourmaline (Fig. 7c). Late SPy pyrite and fine-grained SC chlorite (Fig. 7d) are observed close to the unconformity, with the latter partially replaced along microfractures by fine-grained post hydrothermal alteration SK2 kaolinite (Fig. 7d).

Crystal Chemistry of the Alteration Minerals

Pre-ore alteration C1 chlorites are Mg-Fe-Al chlorite with a calculated structural formula of $\text{Mg}_{5.92}\text{Fe}_{4.20}\text{Al}_{2.54}(\text{Si}_{5.92}, \text{Al}_{2.08})\text{O}_{20}(\text{OH})_{16}$ (Table 1), reflecting the composition

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4 of a clinochlore chlorite. This chemical composition is consistent with formation
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6 temperatures near 275°C based on site occupancy (Cathelineau, 1988). Pre-ore alteration
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8 C2 chlorites have distinctly different chemical compositions than C1 chlorite, with a
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10 calculated structural formula of $\text{Mg}_{2.78}\text{Fe}_{6.65}\text{Al}_{2.36}(\text{Si}_{5.73}, \text{Al}_{2.27})\text{O}_{20}(\text{OH})_{16}$ (Table 1),
11
12 corresponding to the composition of a chamoisite chlorite with formation temperatures
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14 near 300°C. Post-ore chlorites C3 have a calculated average chemical formula of
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16 $\text{K}_{0.08}\text{Mg}_{3.45}\text{Fe}_{1.16}\text{Al}_{5.06}(\text{Si}_{6.61}, \text{Al}_{1.39})\text{O}_{20}(\text{OH})_{16}$ (Table 1), corresponding to the
17
18 composition of a sudoite chlorite with formation temperatures near 185°C. Post-ore
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20 chlorites C4 are similar in composition to C3 chlorite with a calculated structural formula
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22 of $\text{Mg}_{2.80}\text{Fe}_{0.22}\text{Al}_{6.19}(\text{Si}_{6.48}, \text{Al}_{1.52})\text{O}_{20}(\text{OH})_{16}$ (Table 1), corresponding to the composition
23
24 of a sudoite chlorite with formation temperatures near 175°C. Late hydrothermal
25
26 alteration SC chlorites in the Manitou Falls Formation are similar in composition to C3
27
28 and C4 chlorites in the basement with a calculated structural formula of
29
30 $\text{Mg}_{3.09}\text{Fe}_{0.09}\text{Al}_{6.23}(\text{Si}_{6.82}, \text{Al}_{1.18})\text{O}_{20}(\text{OH})_{16}$ (Table 1), corresponding to a composition of a
31
32 sudoite chlorite with formation temperatures near 155°C.

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35 When plotted on a molar Al-Mg-Fe ternary diagram (Fig. 8), C1 and C2 chlorites
36
37 plot on the Fe-Mg side of the diagram. These compositions are similar to those of biotite
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39 in the basement rocks and probably reflect an inherited chemistry from the altered
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41 minerals or alteration fluids that equilibrated with the metamorphic basement rocks prior
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43 to chlorite alteration. The C1 chlorite has a similar composition to pre-ore chlorites
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45 reported by Alexandre et al. (2005) for Dawn Lake, McArthur River and Rabbit Lake
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47 basement deposits whereas C2 chlorites have higher Fe contents (Fig. 8). Such
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49 differences in the chemical composition between the C1 and C2 chlorites might be
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4 attributed to the different chemistry of the altered minerals that locally fixed the
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6 chemistry of the alteration fluid. C3, C4 and SC chlorites plot close to the post-ore
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8 chlorite field from Dawn Lake, McArthur River and Rabbit Lake basement deposits
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10 (Alexandre et al., 2005).
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14 Pre-ore M1 muscovites have similar crystal chemistry to syn-ore M2 muscovite.
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16 Calculated structural formula for pre-ore M1 muscovites is
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18 $K_{0.80}Mg_{0.17}Fe_{0.03}Al_{1.82}(Si_{3.27},Al_{0.73})O_{10}(OH)_2$ with K varying from 0.72 to 0.89, and is
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20 $K_{0.75}Mg_{0.17}Fe_{0.03}Al_{1.83}(Si_{3.29},Al_{0.71})O_{10}(OH)_2$ for syn-ore muscovite, with K varying from
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22 0.65 to 0.89 (Table 1). Temperatures derived from the crystal chemistry of M1 and M2
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24 muscovites are similar at 250°C (Table 1). The crystal chemistry of the pre-ore and syn-
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26 ore muscovites are analogous to those reported by Alexandre et al. (2005) for pre-ore and
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28 syn-ore illites at Dawn Lake, McArthur River and Rabbit Lake, suggesting that the fluids
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30 responsible for the alteration of the basement rocks that formed the pre- and syn-ore
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32 muscovites were consistent throughout the eastern part of the Athabasca Basin. Early
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34 hydrothermal alteration SM muscovites in the Manitou Falls Formation sandstones have
35
36 a similar crystal chemistry to the pre- and syn-ore muscovite (M1, M2) in the basement,
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38 with calculated structural formulas of $K_{0.80}Mg_{0.12}Fe_{0.04}Al_{1.86}(Si_{3.26},Al_{0.74})O_{10}(OH)_2$ with K
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40 varying from 0.71 to 0.85 and formation temperatures near 240°C (Table 1).
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48 XRD data from pre-ore M1 and syn-ore M2 muscovites in the basement and SM
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50 muscovite from the Manitou Falls Formation sandstones indicate the presence of 2
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52 polytypes within the Millennium alteration system (Fig. 9). SM muscovites from the
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54 Manitou Falls Formation sandstones and pre-ore basement M1 muscovites up to 25
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56 meters below the unconformity are dominated by the 1Mt polytype (Figs. 6b, c, 9), while
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pre-ore M1 and syn-ore M2 muscovites deeper than 25 meters below the unconformity are dominated by the 1Mc polytype (Figs. 6d, 9). 1Mc muscovites are rigid micrometer-scale lath-like shaped crystals (Fig. 6d), whereas 1Mt muscovites are thin, sub-micrometer, “hairy-shaped” crystals growing on the edge of 1Mc muscovites (Fig. 6 b, c). These polytypes are similar to those observed in the Manitou Falls Formation at the Shea Creek deposit by Laverret et al. (2006), where the 1Mc polytype was dominant in the lower part of the sandstones and the 1Mt polytype was dominant close to the sandstone/basement unconformity and around brittle structural features. Laverret et al. (2006) interpreted the 1Mc polytype in barren areas to be related to muscovite alteration from diagenetic brines and the 1Mc and 1Mt polytype in the mineralized area to be related to the reinjection of basinal brines that interacted with the basement rocks into the overlying sandstone. At Millennium, the 1Mc muscovite polytype is the dominant alteration mineral in the basement rocks, which we interpret to be related to pre- and syn-ore muscovite alteration from diagenetic brines. The 1Mt muscovite polytype is observed growing from the edge of the 1Mc muscovites, as is the case at Shea Creek, which would imply that this 1Mt polytype is related to late post-mineralization alteration fluids.

Syn-ore basement APS minerals (APS) vary from 9.50 to 10.66% SrO, 1.95 to 2.20% CaO, 2.70 to 3.84% La₂O₃, 4.37 to 6.02% Ce₂O₃, 1.16 to 1.56% Nd₂O₃, 4.70 to 10.29% SO₃, 22.44 to 24.72 P₂O₅, 0.11 to 4.27% FeO and 29.22 to 34.56% Al₂O₃ with minor amount of F, BaO and ThO₂ (Table 2). The calculated average structural formula is Sr_{0.45}Ca_{0.17}LREE_{0.30}(Al_{2.87}Fe_{0.11})(PO₄)_{1.54}(SO₄)_{0.35}(OH)₆, corresponding to goyazite composition. Early hydrothermal alteration basin APS mineral (SAPS) chemical compositions are similar to syn-ore APS chemistry in the basement rocks at

$\text{Sr}_{0.44}\text{Ca}_{0.18}\text{LREE}_{0.28}(\text{Al}_{2.91}\text{Fe}_{0.11})(\text{PO}_4)_{1.53}(\text{SO}_4)_{0.29}(\text{OH})_6$ (Table 2), also corresponding to a goyazite composition.

The concentration of uranium in uraninite varies from 50.2 to 77.4% U (Table 3) and Pb contents also vary, ranging from 0.2 to 23.1% Pb. Other elements present within the uraninite structure include Ca (0.37 to 3.62%), Si (0.10 to 4.74%), Fe (0.09 to 4.74%), Ti (0.03 to 0.19%), V (0.03 to 0.75%) and O (11.58 to 16.70%) (Table 3), consistent with micron-scale becquerellite and coffinite alteration after uraninite.

Isotopic Compositions of the Minerals and Fluids Involved in the Deposit

Stable isotopic compositions were determined for basement pre-ore M1 and syn-ore M2 muscovites, pre-ore C2 and post-ore C3 chlorite, basin early hydrothermal alteration SM muscovite and late hydrothermal alteration SC chlorite (Table 4 and Figure 10).

Basement pre-ore M1 muscovite and C2 chlorite

Measured $\delta^{18}\text{O}$ and δD values of nine pre-ore M1 muscovites vary from +10.2 to +13.7 per mil for $\delta^{18}\text{O}$, and -66 to -52 per mil for δD . These values are similar to those reported in previous studies of the Manitou Falls Formation (Kotzer and Kyser, 1995), but the δD values are distinct from those that produced early muscovite in basement host rocks in the Key Lake (Wilson and Kyser, 1987), McArthur River and Rabbit Lake deposits (Alexandre, 2005), where δD are higher. Using the formation temperature of 250°C for M1 muscovite obtained from the mineral chemistry (Table 1), the calculated

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4 $\delta^{18}\text{O}$ values for the fluid in equilibrium with this muscovite range from +5.3 to +8.9 per
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6 mil, while the δD fluid values range from -36 to -20 per mil (Fig. 10; Table 4).
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9 Measured $\delta^{18}\text{O}$ and δD values of one pre-ore C2 chlorite gives a calculated $\delta^{18}\text{O}$
10 value of +7.0 per mil and δD of -29 per mil for the fluid in equilibrium with this chlorite
11 (Fig. 10; Table 4), similar to pre-ore M1 muscovite in the basement. This fluid value is
12 similar to those reported by Alexandre et al. (2005) for pre-ore chlorite at Dawn Lake,
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14 McArthur River and Rabbit Lake deposits.
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23 *Syn-ore M2 muscovite and uraninite*

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26 Syn-ore M2 muscovites have $\delta^{18}\text{O}$ values from +10.2 to +12.6 per mil, similar to
27 pre-ore M1 muscovite, but lower δD values averaging -96 per mil. Using the formation
28 temperature of 250°C for M2 muscovite obtained from the mineral chemistry (Table 1),
29 the calculated $\delta^{18}\text{O}$ values for the fluid in equilibrium with this muscovite vary between
30 +5.4 and 7.8 per mil. The δD values of the fluids average -104 per mil (Fig. 10; Table 4),
31 and reflect preferential exchange of hydrogen isotopes with more recent fluids moving
32 along fractures in the mineralized zone as fluids with such low values were not present in
33 the basin except recently (Wilson and Kyser, 1987; Kotzer and Kyser, 1995).
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45 Measured $\delta^{18}\text{O}$ values of four uraninite samples vary from -12.9 to +9.8 per mil
46 (Table 4). Using the $\delta^{18}\text{O}$ values and temperatures of the fluids that formed the syn-ore
47 M2 muscovites, pristine uraninites should have $\delta^{18}\text{O}$ values near -4 per mil (Fayek and
48 Kyser, 2000). The large variation in $\delta^{18}\text{O}$ values of the uraninites at Millennium is
49 interpreted to result from variable degrees of uraninite alteration, with high values
50 reflecting alteration during incorporation of Ca, Si and Fe to secondary uranium minerals
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4 whereas low values reflect exchange of the uraninites with relatively recent ^{18}O -depleted
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6 fluids that percolated along faults and fractures (Kotzer and Kyser, 1993b; Fayek and
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8 Kyser, 2000). The low $\delta^{18}\text{O}$ values in uraninites are attributed to recent fluids, as the
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10 Athabasca Basin slowly migrated to higher latitudes since its formation (Ramaekers and
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12 Dunn, 1977), where meteoric waters with lower δD and $\delta^{18}\text{O}$ values are expected.
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19 *Basement post-ore C3 chlorite*

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21 Two post-ore C3 chlorites have $\delta^{18}\text{O}$ values of +8.1 and +11.5 per mil and δD
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23 values of -57 to -55 per mil. Using formation temperatures of 185°C for C3 chlorites
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25 obtained from the mineral chemistry (Table 1), the calculated $\delta^{18}\text{O}$ values for the fluid in
26
27 equilibrium with these chlorites are +5.4 and +8.8 per mil, and the δD fluids values are -
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29 15 and -13 per mil (Fig. 10; Table 4). The δD values are higher than the fluids associated
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31 with the post-ore chlorite at McArthur River (Alexandre, 2005), which range from -80 to
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33 -31 per mil. The values for the fluids that equilibrated with the C3 chlorites also have
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35 higher δD and $\delta^{18}\text{O}$ values than those for the fluids that produced the chlorite alteration in
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37 the gneissic host rock of the Key Lake deposit in the southwest Athabasca basin (Wilson
38
39 and Kyser, 1987), where $\delta^{18}\text{O}_{\text{FLUID}}$ ranged from +0 to +4.1 per mil, and $\delta\text{D}_{\text{FLUID}}$ was from
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41 -61 to -30 per mil.
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51 *Basin SM muscovite and SC chlorite*

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53 SM muscovites from early alteration of the Manitou Falls Formation sandstones
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55 above the deposit have $\delta^{18}\text{O}$ values of +7.0 and 11.3 per mil and δD values of -61 and -50
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57 per mil. These values are similar to pre- and syn-ore muscovites in the basement and to
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4 reported values in previous studies of the Manitou Falls Formation (Kotzer and Kyser,
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6 1995). Using formation temperatures of 240°C for SM muscovite obtained from the
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8 mineral chemistry (Table 1), the calculated $\delta^{18}\text{O}$ and δD values for the fluid that formed
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10 SM muscovite is +1.8 and 6.1 per mil for $\delta^{18}\text{O}$ and -31 and -20 per mil for δD (Fig. 10).
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14 SC chlorites from late alteration of the Manitou Falls Formation sandstones above
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16 the deposit have $\delta^{18}\text{O}$ values between +9.3 and +15.9 per mil and δD values of -64 and -
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18 48 per mil. Using formation temperatures of 155°C for SC chlorite obtained from the
19
20 mineral chemistry (Table 1), the calculated $\delta^{18}\text{O}$ and δD values for the fluid that formed
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22 SC chlorite are +5.5 and +12.1 per mil for $\delta^{18}\text{O}$ and -20 and -4 per mil for δD (Fig.10).
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30 **Timing of Fluids that have Affected the Millennium Deposit**

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32 Chemical U-Pb ages derived from the composition of uraninite vary from 67 to
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34 702 Ma, suggesting variable recrystallization and Pb loss. There is a decrease in the
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36 chemical age with an increase in Ca, Fe and Si contents of the uraninites (Fig. 11) and
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38 since alteration is expected to result in a decrease in radiogenic Pb, the age of formation
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40 of the uraninite in the Millennium deposit can be estimated by extrapolating the chemical
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42 U-Pb ages to the age when the concentration of the secondary elements (Ca, Fe and Si)
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44 were negligible in the uraninite (Alexandre et al., 2005). Regression to zero content of
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46 these elements in the uraninites corresponds to 1590 Ma (Fig. 11), which is the same as
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48 the initial formation age that Alexandre et al. (2009) proposed for all the unconformity-
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50 related deposits in the Athabasca Basin.
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56 U-Pb isotopic compositions of several uraninite grains analyzed by LA-HR-ICP-
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58 MS and LA-HR-MC-ICP-MS (Table 5) show significant resetting of U-Pb isotopic
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systems (Fig. 12a). This resetting is also evident from the finely mottled and speckled appearance of uraninite grains under reflected light and back scattered electron imaging (Fig. 5f). Almost all uraninite grains have near-zero-age lower intercepts (Fig. 12a), reflecting major recent Pb loss. The $^{206}\text{Pb}/^{207}\text{Pb}$ dates range between 131 ± 30 Ma and 1421 ± 7 Ma (Fig. 12f), $^{206}\text{Pb}/^{238}\text{U}$ dates range between 166 ± 7 Ma and 1212 ± 102 Ma and $^{207}\text{Pb}/^{235}\text{U}$ dates range between 263 ± 9 Ma and 1135 ± 1 Ma (1σ) (Table 5).

U-Pb upper intercept correspond to dates of 1344 ± 48 Ma, 1278 ± 200 Ma, 1091 ± 26 Ma and 804 ± 120 Ma and a lower intercept near 0 Ma (Fig. 12b-e), which are similar to the dominant Pb-Pb dates recorded by the uraninites at Millennium (Fig. 12f). Similar dates for these upper intercept with the most frequent Pb-Pb isotope dates indicate episodes of partial resetting that occurred after the main mineralization event, the most dominant being near 1100 Ma (Fig. 12d) and coincident with the Grenville Orogeny (Hoffman 1990; Mosher 1996).

Pre-ore muscovite samples from the basement have disturbed $^{40}\text{Ar}/^{39}\text{Ar}$ spectra with pseudo-plateau $^{40}\text{Ar}/^{39}\text{Ar}$ dates of 1425 ± 8 Ma and 1435 ± 8 Ma from 41% of the gas released (Fig. 13a). Both samples analyzed show a limited amount of superficial $^{40}\text{Ar}^*$ loss (Fig. 13a) that should not affect the dates. A cumulative probability diagram for the two samples indicates a main date near 1430 Ma, defined by 60% of the steps and corresponding to the pseudo-plateau ages, and a secondary date defined by 15% of the steps near 1390 Ma (Fig. 13b).

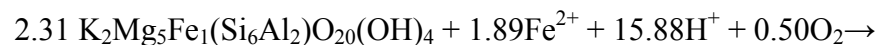
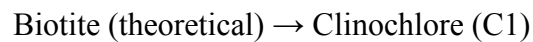
Chemistry and Geochronology of the Leachable Phases

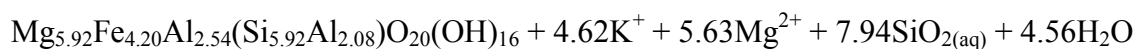
Holk et al. (2003) demonstrated that the isotopic composition of leachable Pb could be used to trace Pb loss events from uranium deposits. The isotopic compositions of Pb leach from around the Millennium deposit indicate the presence of radiogenic Pb ($^{206}\text{Pb}/^{204}\text{Pb} > 30$) (Table 6) in the basement up to 75m above and more than 100m below the mineralized zone (Fig. 14a). The Pb leached from the basin gives a model $^{207}\text{Pb}/^{206}\text{Pb}$ isochron age of 2875 ± 360 Ma, whereas basement rocks give a model $^{207}\text{Pb}/^{206}\text{Pb}$ isochron age of 1075 ± 400 Ma, similar to the 1091 Ma major resetting event recorded by the uraninite.

Discussion

Reactions during alteration and mineralization

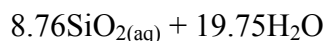
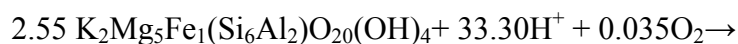
Petrographic relationships reveal four principal phases of alteration: a pre-ore stage with minor C1 clinocllore alteration forming prior to pervasive M1 muscovite alteration, a pre-ore stage with minor C2 chamoisite altering M1 muscovite, a syn-ore stage with coeval precipitation of uraninite, M2 muscovite and H1 hematite and a post-ore stage where sudoite (C3 and C4) is the dominant alteration mineral. Based on petrographic observations, the starting mineral assemblage in the basement rocks is comprised on average of 40% quartz, 25% plagioclase and 20% biotite with minor amounts of potassium feldspar, cordierite, muscovite, graphite, hematite, rutile, pyrite and amphibole. We estimate that the first chloritization event altered about 50% of the biotite to C1 clinocllore according to the reaction:



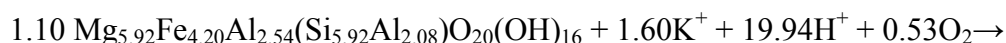


This reaction liberated K, Mg and $\text{SiO}_{2(\text{aq})}$ while consuming Fe, H^{+} and O_2 , creating about 485Å (41%) of space per 2.31 biotite (497.79Å per formula unit) altered to clinochlore (699.59Å per formula unit), for a total space created in the rock of 4% during this phase of alteration. The C1 clinochlore alteration was followed by M1 muscovite alteration of remaining biotite, C1 clinochlore and plagioclase according to the following reactions:

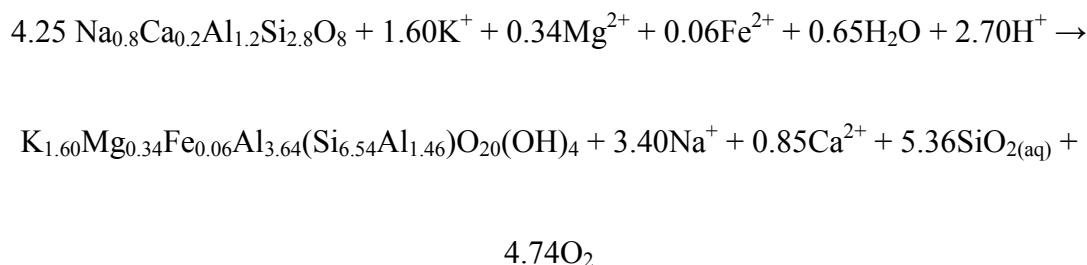
Biotite (theoretical) → M1 Muscovite



C1 Clinochlore (as probed) → M1 Muscovite (as probed)

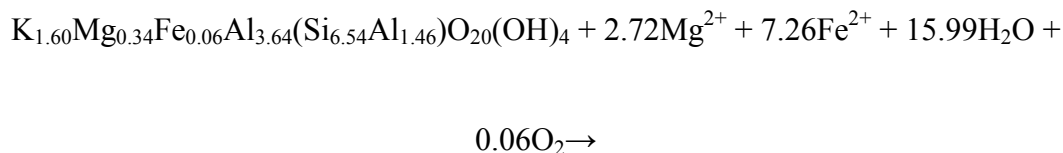
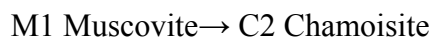


Oligoclase (theoretical) → M1 Muscovite (as probed)



Based on petrographic observations, we estimate that both biotite and clinocllore alteration to muscovite accounted for 22.5% each of the muscovite created while oligoclase alteration accounted for 55%. These reactions liberated Mg, Fe, Na, Ca, $\text{SiO}_{2(\text{aq})}$ and O_2 , while consuming K and H^+ . Up to 65% of the K consumed during these reactions can be derived from biotite, whereas the rest of the K had to be brought in with the fluids responsible for M1 muscovite alteration. This muscovite alteration (467.665 Å per formula unit) created 802 Å (63%) of space per 2.55 formula unit of altered biotite, 281 Å (37%) of space per 1.10 formula unit of altered C1 clinocllore and 203 Å (30%) of space per 4.25 formula units of altered oligoclase (158.26 Å per formula unit). The total space created in the rock during this alteration is 21%, bringing the total space created in the rock during both pre-ore alteration events to ca. 25% if no muscovite was precipitated in the void created during the M1 muscovite basement alteration phase.

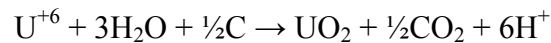
The M1 muscovite alteration was followed by minor C2 chamoisite alteration of M1 muscovite according to the following reaction:



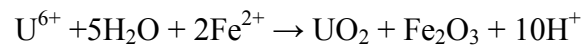


This reaction liberated K, $\text{SiO}_{2(\text{aq})}$ and H^{+} while consuming Mg, Fe and O_2 . The elements consumed during this reaction could have been derived from the alteration of basement rocks during pre-ore M1 muscovite alteration. This C2 chamoisite alteration used 308Å (66%) more space per 1.10 formula unit (704.40Å per formula unit) than the muscovite, consuming voids created during the earlier alteration phases. The total space consumed during this phase was small as chamoisite alteration is very localized. However, this alteration locally reduced the porosity of the system by partially filling some of the voids created during the earlier alteration phases.

The mineralizing event was characterized by the precipitation of uraninite, from the same fluid that precipitated M1 muscovite, and by formation of euhedral M2 muscovite and H1 hematite in the voids created during the pre-ore stage. The uraninite precipitated by reduction of the basement where fluids encountered graphitic and Fe-rich chlorite lithologic units, following the reaction:



Hematite was also precipitated simultaneously with uraninite following the reaction:



Syn-ore M2 muscovite precipitation required K, Fe, Al and $\text{SiO}_{2(\text{aq})}$ while H1 hematite required Fe. The Al, SiO_2 and up to 6% of the Fe consumed during these reactions can be derived from the previous alteration stages, whereas the rest of the Fe

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4 and the totality of the K had to be brought in with the fluid responsible for the syn-ore
5 alteration.
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9 The post mineralization fluid would have been enriched in Ca, Mg and Na.
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11 Derome et al. (2005) also observed a Ca-rich brine in their fluid inclusion study at
12
13 McArthur River, which they suggested was derived from a Na-rich basinal brine that
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15 infiltrated and reacted with the basement rocks, resulting in a Ca-rich brine. The main
16
17 post-ore alteration minerals consist of dravite, calcite and sudoites (C3, C4), which
18
19 precipitated in the space created during the pre-ore alteration stage from the chemically
20
21 modified fluids, considerably reducing the porosity and sealing the system to subsequent
22
23 fluid infiltration in the basement rocks, except along fractures. Boron for the dravite
24
25 might have been derived from the alteration of tourmaline in granitoids in the basement
26
27 rocks or in detrital tourmaline in the Manitou Falls Formation sandstones, while the Ca
28
29 needed for calcite was available in the chemically modified post-ore residual fluid. The
30
31 residual elements after dravite and calcite post-ore alteration phases would be Mg, Al and
32
33 Si, crystallizing late sudoite (C3 and C4).
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41 Alexandre et al., (2005) also found similar results at the Dawn Lake, McArthur
42
43 River and Rabbit Lake deposits, where pre-ore illite alteration of the basement rocks
44
45 created space while liberating the Fe and Mg needed for pre-ore chlorite and hematite.
46
47 The chlorite alteration consumed space, reducing the porosity and closing the system.
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49 However, the Mg liberated by alteration of the basement rocks was principally used to
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51 produce pre-ore clinochlore at those deposits, whereas post-ore sudoite consumed Mg at
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53 the Millennium deposit.
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4 *Origin of the fluids responsible for alteration and mineralization of the Millennium*
5
6 *deposit*
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9 Pre-ore alteration of the basement rocks at Millennium consists of minor amounts
10 of C1 clinocllore on the outside edge of the alteration halo followed by an extensive M1
11 muscovite alteration throughout the altered area. A formation temperature of 275°C for
12 C1 chlorite, that pre dates the pre-ore M1 muscovite alteration, suggests that C1 chlorite
13 alteration of biotite may reflect retrograde metamorphism in the basement rocks
14 following peak of the Trans-Hudson Orogeny prior to deposition of the Athabasca Basin.
15 The temperature of formation and the isotopic compositions of the fluids in equilibrium
16 with pre-ore M1 muscovite are similar to those of SM muscovite in the sandstone above
17 the Millennium deposit (Fig. 10), suggesting the same fluids were responsible for the
18 alteration of both Manitou Falls Formation sandstones and underlying basement rocks.
19 The Ca-Sr-rich and LREE-Th-poor nature of the APS mineral found in the Manitou Falls
20 Formation sandstones is consistent with leaching of detrital apatite in the Manitou Falls
21 Formation, suggesting a basin origin for the fluids responsible for the early hydrothermal
22 alteration and pre-ore stage. Pre-ore M1 and early hydrothermal alteration SM muscovite
23 alteration is therefore associated with a Na-K-Fe basin-derived fluid crystallizing the 1M_c
24 polytype at temperatures near 250°C. This muscovite alteration created the full extent of
25 the alteration halo present at Millennium while creating voids used during later alteration
26 phases, including that associated with the uranium mineralization.
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53 Minor amounts of C2 chamoisite altered the M1 muscovite in the footwall of the
54 fault associated with the cordierite-graphite metapelitic gneiss unit. The fluid that
55 precipitated C2 chamoisite has stable isotopic compositions identical to those of the
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basinal brine that precipitated M1 muscovite, but the Mg needed to precipitate C2 chamoisite has to originate from alteration of basement rocks by the basinal brine, as the basinal brine was Mg-poor. The reaction from pre-ore muscovite to pre-ore chamoisite consumed O₂, also suggesting a precipitation from oxidized basinal fluids. Therefore, chemically modified basinal fluids were created from reaction between the basinal brine and basement rocks and resulted in Mg-Fe-rich fluids that were responsible for the small volume of C2 chamoisite in the footwall of the cordierite-graphite metapelitic gneiss unit.

Syn-ore alteration consists mainly of M2 muscovite and minor APS minerals. Calculated $\delta^{18}\text{O}_{\text{FLUID}}$ values for syn-ore muscovites are similar to those of pre-ore muscovites, whereas the calculated $\delta\text{D}_{\text{FLUID}}$ for syn-ore muscovites are significantly lower than pre-ore muscovites (Fig. 10). Low $\delta\text{D}_{\text{FLUID}}$ values have been observed at Key Lake and McArthur River for alteration minerals both near and distal from the ore and reflect preferential hydrogen isotope exchange with later high latitude meteoric fluids that was facilitated by fractures and faults (Wilson and Kyser, 1987; Kotzer and Kyser, 1991; Kotzer and Kyser, 1995). Even given this later exchange, the stable isotopic compositions of the fluid in equilibrium with syn-ore muscovites approach those of pre-ore muscovites. Moreover, syn-ore APS mineral compositions are similar to the early hydrothermal alteration substage APS minerals in the Manitou Falls Formation, suggesting similar formation fluids for those two phases, and reinforcing a basin origin for the fluid responsible for the mineralization. Syn-ore alteration is therefore associated with precipitation of uraninite, muscovite and iron oxide in the footwall of the fault related to the cordierite-graphite metapelitic gneiss from a basin-derived fluid crystallizing the 1Mc muscovite polytype at temperatures near 250°C.

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4 Post-ore alteration is characterized by dravite veins, calcite veins and pervasive
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6 sudoite crystallization (C3, C4) filling the voids remaining from pre- and syn-ore stage.
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8 Calculated δD_{FLUID} and $\delta^{18}O_{\text{FLUID}}$ values for post-ore C3 chlorite are similar to the basinal
9
10 fluids that precipitated M1 muscovite (Fig. 10; Table 4). Chemically, as for C2
11
12 chamoisite, the basinal fluids required Mg to precipitate C3 and C4 sudoites, which was
13
14 acquired from the alteration of basement rocks. The isotopic composition and lower
15
16 formation temperature of SC sudoite in the Manitou Falls Formation sandstones are
17
18 similar to those of post-ore C3 and C4 chlorite in the basement. The similarity in
19
20 chemical and isotopic composition between the sandstone and basement sudoites
21
22 suggests that post-ore C3 and C4 sudoite is likely coeval with late hydrothermal
23
24 alteration sudoite (SC). Post-ore alteration is therefore associated with Ca-Mg chemically
25
26 modified basinal fluids that precipitated sudoite at temperatures near 185°C, indicating a
27
28 decrease in temperature of ca. 65°C since the syn-ore alteration stage. Precipitation of
29
30 sudoite in the basement used the remaining voids created during the pre-ore stage by
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32 infiltration of basinal fluid, reducing the porosity of the system.
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43 *Timing of mineralization and alteration*

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45 The oldest date obtained at Millennium is 1590 Ma extrapolated from the
46
47 chemical age of unaltered uraninite. This possible mineralization age is identical to the
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49 age reported by Alexandre et al. (2009) for sandstone- and basement-hosted deposits
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51 throughout the Athabasca Basin and is consistent with the hypothesis that the major,
52
53 primary mineralizing event in all the Athabasca Basin deposits occurred at about 1590
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55 Ma.
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Wilson et al. (1987) demonstrated that the interlayer region of fine-grained muscovite can interact with residual basinal brines and late meteoric waters at low temperatures lowering the K-Ar ages. The ca. 1430 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ date for the pre-ore muscovite are older than the major resetting ages obtained from U-Pb on uraninite, but younger than the age of primary mineralization at 1590 Ma. This suggests that the muscovite was altered during subsequent fluid circulation events or that there was residual fluid in the basement reacting with the muscovite until 160 m.y. after the primary mineralization event.

A maximum age of post-ore C3 sudoite can be estimated from the leachable Pb data if we assume that all the radiogenic Pb present in post-ore C3 sudoite samples resulted from decay of the uranium mineralization and was leached from the mineralization during the C3 sudoite alteration phase. If the C3 sudoite alteration occurred shortly after the mineralizing event, only a small amount of radiogenic Pb would have had the time to form in the deposit. The typical $^{206}\text{Pb}/^{204}\text{Pb}$ ratio in basement rocks affected by C3 sudoite is 50, which can be used to constrain how much time passed after the uranium ore formed to produce enough ^{206}Pb to elevate the $^{206}\text{Pb}/^{204}\text{Pb}$ ratio in the basement to 50. Based on a volume of $5 \times 10^6 \text{ m}^3$ of basement rocks affected by the fluid that precipitated the post-ore C3 sudoite, and assuming that all the radiogenic ^{206}Pb available within the mineralization was leached from the deposit at the time of C3 sudoite crystallization, the minimum amount of time required for the uranium to decay to enough ^{206}Pb would be 100 m.y. after the ore forming event. This would give an estimated age of ca. 1490 Ma for the C3 sudoite event, similar to the 1430 Ma date obtained from $^{40}\text{Ar}/^{39}\text{Ar}$ on pre-ore alteration M1 muscovite. Given that the date from the M1 muscovite

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4 is the time when fluids were last in the basement interacting with muscovite, a maximum
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6 age for the last alteration event recorded by both the $^{40}\text{Ar}/^{39}\text{Ar}$ systematic of the M1
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8 muscovite and the C3 sudoite is 1430 Ma, 160 m.y. after the mineralization event.
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11 Several episodes of resetting of the U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ isotope systematics are
12
13 recorded in uraninite grains at ca. 1390, 1350, 1275, 1090 and 804 Ma (Fig. 12a-f),
14
15 reflecting episodic Pb loss from the deposit. These resetting events correspond to fluid
16
17 circulation in fractures in the Athabasca Basin induced by far-field tectonic events. The
18
19 1390 Ma is similar to the ca. 1400 Ma resetting age obtained by Alexandre et al. (2009)
20
21 on pre-ore and syn-ore illites and chlorites at Dawn Lake, McArthur River and Rabbit
22
23 Lake basement-hosted deposits and the Virgin River prospect, which they interpreted as
24
25 fluid movement in the Athabasca Basin induced by the distal Bertoud Orogeny (Nyman
26
27 et al., 1994; Sims and Stein, 2001), disturbing the $^{40}\text{Ar}/^{39}\text{Ar}$ systematics in muscovites
28
29 from the basement. The ca. 1350 Ma partial resetting event corresponds to the 1360-1370
30
31 Ma regional rifting and deformation associated with the East Kootenay Orogeny
32
33 southwest of the Athabasca Basin (McMechan and Price, 1982; Doughty and
34
35 Chamberlain, 1996; Beaudoin, 1997). The ca. 1275 Ma date is interpreted to correspond
36
37 to a well-known fluid circulation event in the Athabasca Basin induced by the
38
39 emplacement at ca. 1270 Ma of the MacKenzie Dike swarms (LeCheminant and Heaman,
40
41 1989). The ca. 1090 Ma date is not commonly recorded elsewhere in deposits of the
42
43 Athabasca Basin as a dominant age and does correspond to the timing of the Grenville
44
45 Orogeny (Hoffman 1990; Mosher 1996). Millennium is located in the Wollaston-
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47 Mudjatik transition zone whereas other deposits are hosted within the Wollaston Domain,
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49 so that the former setting may have facilitated resetting of Millennium ores during the
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4 Grenville Orogeny because there was fluid movement in the transition zone. The $804 \pm$
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7 120 Ma date to fluid circulation due to the break up of Rodinia (Mayers et al., 1996;
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9 Condie, 2001).

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14 *Comparison of the alteration zone with other basement deposits*

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16 The alteration zone present at Millennium consists of a weak distal retrograde
17
18 metamorphism C1 clinocllore alteration (>75 meters from the ore zone), a pervasive
19
20 proximal pre-ore M1 muscovite alteration (0 to 100 meters from the ore zone) and a minor
21
22 zone of pre-ore C2 chamoisite in the footwall of the fault associated with the cordierite-
23
24 graphite metapelitic gneiss unit. Uranium mineralization is closely associated with syn-
25
26 ore M2 muscovite and H1 hematite. During the post-ore stage, pre- and syn-ore minerals
27
28 were overprinted by dravite and sudoite. In contrast, alteration zones in basement-hosted
29
30 deposits at Dawn Lake, McArthur River and Rabbit Lake consist of a weaker distal pre-
31
32 ore muscovite alteration zone (30-100 meters from the ore zone) and a stronger proximal
33
34 pre-ore clinocllore alteration zone (0 to 30 meters from the ore zone). Mineralization in
35
36 all basement-hosted deposits is closely associated with syn-ore euhedral muscovite and
37
38 hematite while post-ore alteration consists of sperulitic dravite in voids, calcite and
39
40 clinocllore veinlets (Alexandre et al, 2005).

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48 Isotopic compositions of pre-ore muscovite from Millennium, Dawn Lake,
49
50 McArthur River and Rabbit Lake deposits (Alexandre et al, 2005) all reflect a consistent
51
52 230-250°C basinal brine for their origin. Pre-ore C2 chamoisite at Millennium is related
53
54 to the presence of a chemically modified basinal fluids produced by interaction of the
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56 basinal fluid with basement rocks, as it is the case at Dawn Lake, McArthur River and
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4 Rabbit Lake (Alexandre et al., 2005). The observed temperatures of these basement-
5
6 related fluids range from 180 to 300°C, and the higher temperature observed at
7
8 Millennium and McArthur River may reflect the depth of the chemically modified basinal
9
10 fluids at Millennium and McArthur compared to the fluid at the Dawn Lake and Rabbit
11
12 Lake deposits.
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15
16 The mineralizing event at the Millennium, Dawn Lake, McArthur River and
17
18 Rabbit Lake deposits is associated with syn-ore muscovite and hematite at a temperature
19
20 around 250°C. The isotopic compositions of syn-ore muscovites are similar among
21
22 Millennium, Dawn Lake, McArthur River and Rabbit Lake deposits, and are consistent
23
24 with a basin origin for the fluids that precipitated uranium in all the basement-hosted
25
26 deposits in the Athabasca Basin. The same minerals forming during the syn-ore stage
27
28 combined with similar fluid isotopic compositions, chemical compositions and formation
29
30 ages for Millennium and all other basement deposits support the model proposed by
31
32 Alexandre et al., (2009) wherein unconformity-related uranium deposits in the Athabasca
33
34 Basin formed from basinal brines at around the same time in response to a far field
35
36 tectonic event.
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43 The isotopic composition and chemistry of post-ore sudoite (C3 and C4) suggests
44
45 chemically modified basinal fluids derived from the alteration of basement rocks by a
46
47 basinal brine at Millennium, as it is the case at McArthur River (Alexandre et al., 2005).
48
49 The formation temperature at the Millennium and Rabbit Lake deposits also differs, with
50
51 185°C for C3 and C4 sudoite at Millennium compared to 135°C for post-ore clinochlore
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53 at Rabbit Lake. The formation temperature of post-ore sudoite at Millennium is similar to
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55 pre-ore chlorite found at Dawn Lake.
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Implications for exploration of Millennium-Type deposits

Leachable radiogenic Pb is found at the Millennium deposit in the basement up to 75m above and more than 100 meters below the mineralized zone (Fig. 14a). The Pb leached from basement rocks gives a model Pb-Pb age of 1075 ± 400 Ma, similar to the 1091 Ma major resetting event recorded by the uraninite. This indicates that major Pb loss from the ores occurred during the Grenville orogeny, concordant with what was recorded in the U-Pb isotope systematics in the uraninites. Radiogenic Pb is also locally observed near the unconformity (Fig. 14a, b, c), which is expected as the basal Manitou Falls Formation has been a conduit for fluids throughout the history of the basin (Holk et al, 2003). In contrast to what is observed above sandstone-hosted deposits (Holk et al, 2003; Kister et al., 2004), radiogenic Pb is absent in the Manitou Falls Formation sandstones above Millennium (Fig. 14b, c). Leachable Pb from Manitou Falls Formation sandstones above Millennium gives a model Pb-Pb age of 2875Ma, which is older than the Athabasca Basin. Therefore, radiogenic Pb in the sandstone above Millennium does not host radiogenic Pb derived from the Millennium deposit, and has to be derived from an older source, such as the detrital components in the basin fill. The data from Millennium show that the Pb mobilized from the deposit by different resetting events did not infiltrate the sandstone, except along fractures, and affected the basement up to 100 meters from the ore zone. Hence, to apply this technique to find basement deposits, basement units and fractures in the Manitou Falls Formation should be sampled during drill program.

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4 The atypical halo present at the Millennium deposit is reflected by the absence of
5
6 predominant pre-ore clinocllore alteration paragenetically later than pre-ore muscovite, a
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8 greater amount of post-ore dravite and the presence of post-ore sudoite. In terms of fluids,
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10 the atypical alteration zone at Millennium can be attributed to the scarcity of chemically
11
12 modified basinal fluids during the pre-ore alteration phase. The presence of chemically
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14 modified basinal fluids crystallizing clinocllore at Dawn Lake, McArthur River and
15
16 Rabbit Lake or chamoisite at Millennium in the early phases in the evolution of the
17
18 deposits seems to be critical for basement-hosted deposits. The amount of chemically
19
20 modified basinal fluids produced by the system during the early alteration phases may
21
22 produce a more pronounced and effective redox front if the pre-ore clays are ferrous,
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24 resulting in a larger and richer deposit, as it is the case at the McArthur River deposit.
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33 **Conclusions**

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36 The atypical alteration halo at Millennium is the result of pervasive muscovite
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38 alteration of the basement rocks by Na-K-Fe basinal brines during the pre-ore stage at ca.
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40 250°C and the lack of predominant pre-ore clinocllore alteration. As alteration of the
41
42 basement rocks progressed, the basinal brines acquired Ca, Fe and Mg and precipitated
43
44 chamoisite at ca. 300°C at a similar stratigraphic level as the mineralization. This
45
46 chamoisite alteration might have been one of the reductants for the precipitation of
47
48 uraninite during the syn-ore stage. Pre-ore muscovite and chamoisite alteration were
49
50 important for the formation of the deposit as basement rocks altered to muscovite by
51
52 basin-derived fluids created up to 20% voids in the basement rocks while chemically
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54 modified basinal fluids emplaced a Fe-rich chamoisite that demarcates a redox front
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4 during the syn-ore stage where uranium ore was precipitated. Uranium mineralization
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6 occurred at ca. 1590Ma primarily from basin-derived fluids. Post-ore dravite and sudoite
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8 precipitated from the residual Ca-Na-Mg fluids filling the voids created during the pre-
9
10 ore stage and sealing the system. Timing of post-ore sudoite alteration was no more than
11
12 160 m.y. after the mineralizing event. Pb mobilized from the deposit by different
13
14 resetting events did not infiltrate the sandstone, except along fractures, and affected the
15
16 basement up to 100 meters from the ore zone.
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Figure Captions

Figure 1: Simplified geologic map of the Athabasca Basin in northern Saskatchewan, Canada. The major lithotectonic units of its basement (provinces and their domains) are indicated in *italics*. Also shown is the position of the Millennium deposit (star) and major unconformity-related uranium deposits (circle). BLSZ: Black Lake Shear Zone; VRSZ: Virgin River Shear Zone. Modified after Sibbald and Quirt (1987).

Figure 2: (A) Schematic plan view of the orebodies of the basement-hosted Millennium deposit with the location and orientation of the drill holes sampled, (B) and (C) schematic cross-sections of lithology and structure of the Manitou Falls Formation above the basement-hosted Millennium deposit. Also shown is the location of the drill holes sampled and samples taken.

Figure 3: (A) Schematic cross-section of lithology, structure and orebodies and (B) alteration of the basement-hosted Millennium deposit based on detailed logging of three drill cores and compilation of drill cores logged by Cameco Corp. (Modified from Roy et al., 2005). Also shown is the location of the drill holes sampled and samples taken.

Figure 4: General mineral paragenesis for (A) basement rocks and (B) Manitou Falls Formation sandstones of the Millennium deposits. Three main alteration stages for the basement rocks include pre-, syn-, and post ore alteration and four main stages for the Athabasca Group sandstones include early, peak and late diagenesis and late alteration. The thickness of the lines indicates the relative abundance. Dashed lines indicate uncertainty in the position. Also shown is the primary mineral present in pelitic rocks (*), granites and pegmatites (†) and amphibolite (‡). Temperatures associated with the different alteration minerals are derived from their chemistry and are accurate to within 30°C (see text for detailed explanation).

Figure 5: Photomicrographs of typical basement rock mineral phases from pre-, syn- and post-ore. (A) fresh metapelite sample containing biotite, quartz and plagioclase. (B) Pre-ore C1 chlorite altering biotite and rutile (R0) to R1 rutile crystals. (C) Pre-ore M1 muscovite alteration within metapelite. (D) Pre-ore C2 Fe-rich chlorite altering pre-ore M1 muscovite. (E) Massive uraninite (U) and associated syn-ore M2 muscovite and H1 iron oxide. (F) Massive uraninite (U) showing a fine mottled and speckled appearance and associated syn-ore muscovite (M2). (G) Post-ore dravite (T) cross-cutting syn-ore M2 muscovite being altered by C3 chlorite. (H) Post-ore chlorite (C4) altering pre-ore muscovite (M1). PP: Plane polar light; XP: Cross-polar light; RL: Reflected light.

Figure 6: Secondary Electron Scanning Electron Microscope microphotographs of (A) basement uraninite alteration to acicular coffinite (Cof). (B) detrital quartz (SQ0), peak diagenesis muscovite (SM) and peak diagenesis dickite (SK1) of Athabasca Group sandstone with peak diagenesis 1Mc muscovite showing overgrowth of the 1Mt polytype. (C) basement pre-ore muscovite (M1) close to the unconformity having the 1Mt polytype. (D) basement pre-ore muscovite (M1) below the ore zone having the 1Mc polytype.

Figure 7: Photomicrographs of typical Athabasca Group sandstones mineral assemblages from early, peak and late diagenesis and late alteration. (A) SH1 hematite highlighting the boundary between detrital quartz (SQ0) and SQ1 quartz overgrowth. SM muscovite

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4 and SK1 dickite are altering SQ1 quartz overgrowth. (B) Peak diagenesis muscovite
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6 (SM) and dickite (SK1) being cross-cut by dravite (ST). (C) Hematite (SH2) cross-
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8 cutting dravite (ST) and peak diagenesis muscovite (SM) and dickite (SK1). (D) Late
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10 diagenesis chlorite (SC) cross-cut by late kaolinite (SK2).
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18 Figure 8: (A) Al-Mg-Fe triangular diagram for SC (cross) late diagenesis, C1 (diamond)
19 pre-ore, C2 (square) pre-ore, C3 post-ore (circle) and C4 (triangle) post-ore chlorites
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21 from the Millennium deposit, plotted as a function of molar proportions (Bailey, 1980).
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23 The black solid lines delineate the field of pre-ore and post-ore basement-derived chlorite
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25 from Dawn Lake, McArthur River and Rabbit Lake basement-hosted deposits (Alexandre
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27 et al., 2005). Also shown is the composition of biotite, muscovite, clinocllore and
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29 sudoite.
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39 Figure 9: XRD profiles of 2-5 and <2 μm muscovites separates from early hydrothermal
40 alteration (SM) from the Athabasca Basin and pre-ore muscovites (M1) from the
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42 basement of Millennium showing 1Mc polytype as the main polytype in the basement
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44 and 1Mt as the main polytype close to the sandstone-basement unconformity. Major peak
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46 position for 1Mc, 1Mt and 2M₁ muscovite polytypes are also shown (Drits et al., 1993).
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56 Figure 10: Calculated δD and $\delta^{18}\text{O}$ values of fluids in equilibrium with clay minerals
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58 from various alteration stages from basin and basement and $\delta^{18}\text{O}$ values of uraninite at
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the Millennium deposit (Table 4). Also shown are the meteoric water line (MWL) the isotopic composition of standard modern ocean water (SMOW) and the effect of preferential exchange of H isotopes with modern fluids.

Figure 11: Plot of Ca+Fe+Si contents in uraninites (Table 3) as a function of chemical date. Regression to 0% Ca+Fe+Si gives a chemical uraninite age of 1590 Ma.

Figure 12: Concordia diagram for Millennium uraninites showing (A) all the intercepts, (B) a resetting upper intercept at 1344 Ma, (C) a resetting upper intercept at 1278 Ma, (D) a resetting upper intercept at 1091 Ma, (E) a resetting upper intercept at 826 Ma and (F) a cumulative probability histogram of $^{207}\text{Pb}/^{206}\text{Pb}$ apparent-ages in Ma. Data from Table 5. Bold numbers beside analyses represent $^{207}\text{Pb}/^{206}\text{Pb}$ ages.

Figure 13: (A) ^{40}Ar - ^{39}Ar spectra for pre-ore muscovites at Millennium showing pseudo-plateau $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 1425 ± 8 Ma and 1435 ± 8 Ma; (B) histogram of cumulative probability showing major perturbations of the $^{40}\text{Ar}/^{39}\text{Ar}$ systematics at 1430 and 1391 Ma.

Figure 14: Distribution of $^{206}\text{Pb}/^{204}\text{Pb}$ ratios for leachable Pb from the Millennium deposit and surroundings. (A) W-E cross-section in the basement, (B) W-E cross-section in the

Athabasca sandstone across the deposit and (C) N-S cross-section in the Athabasca
sandstone along the deposit.

Figure 01

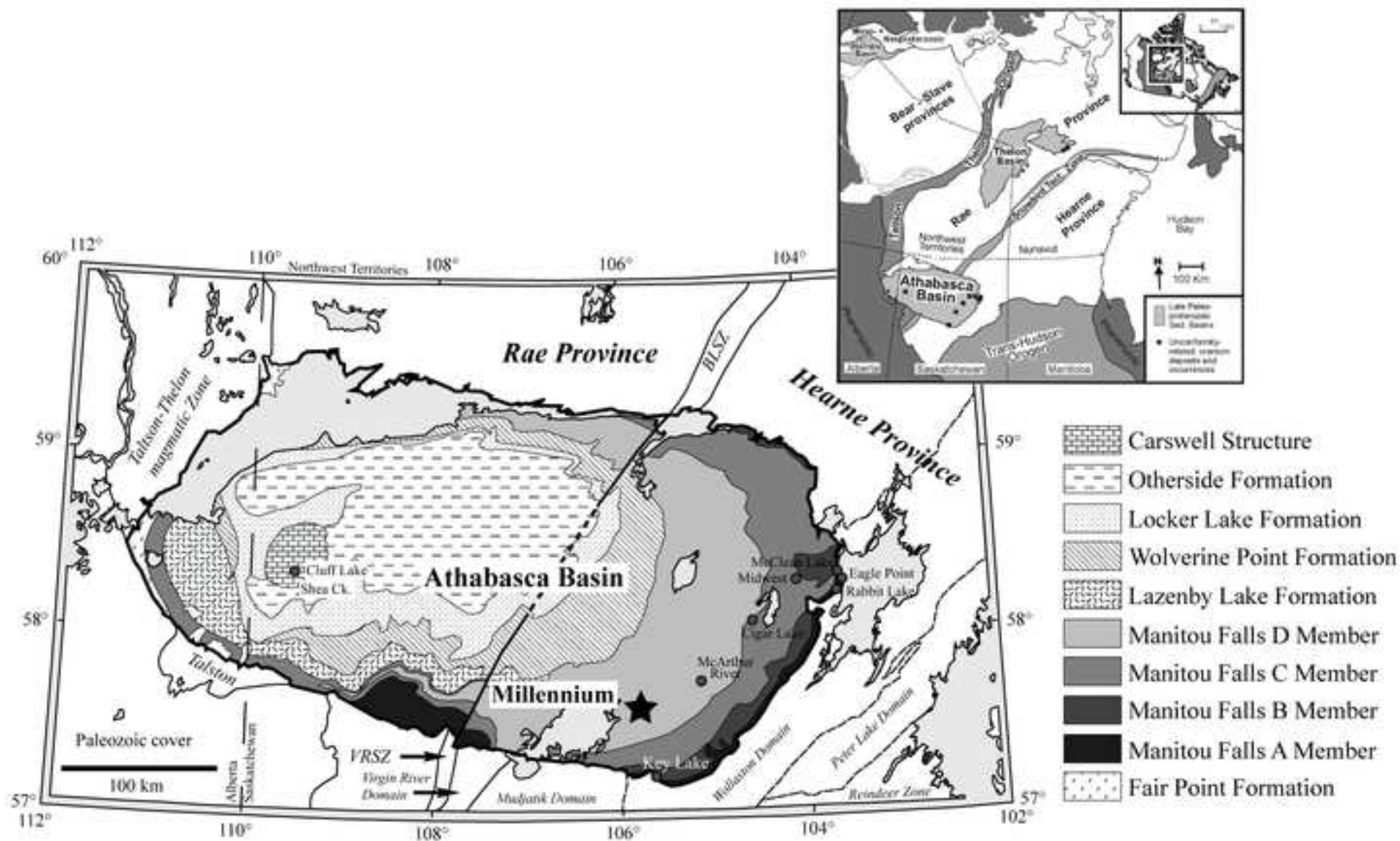
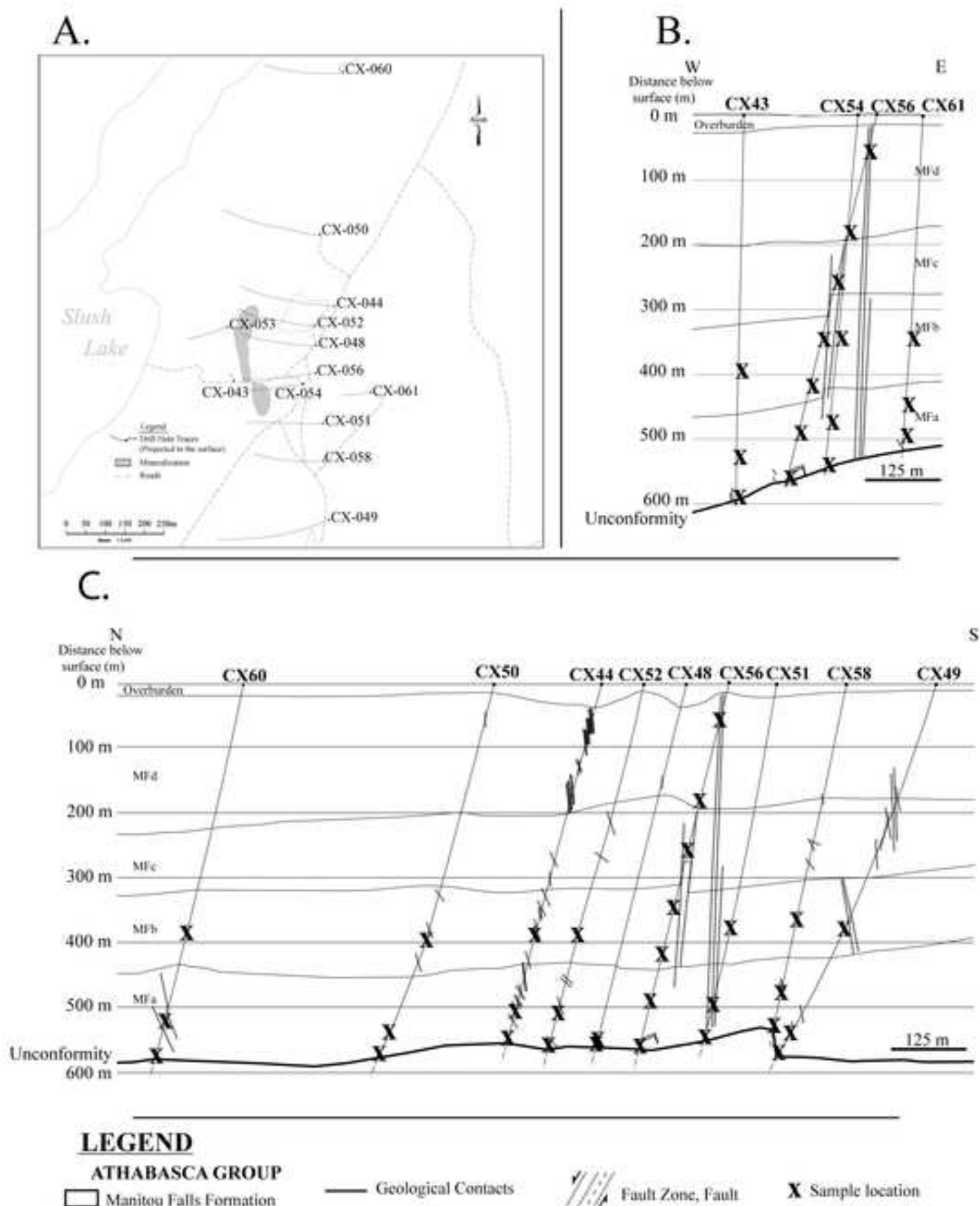


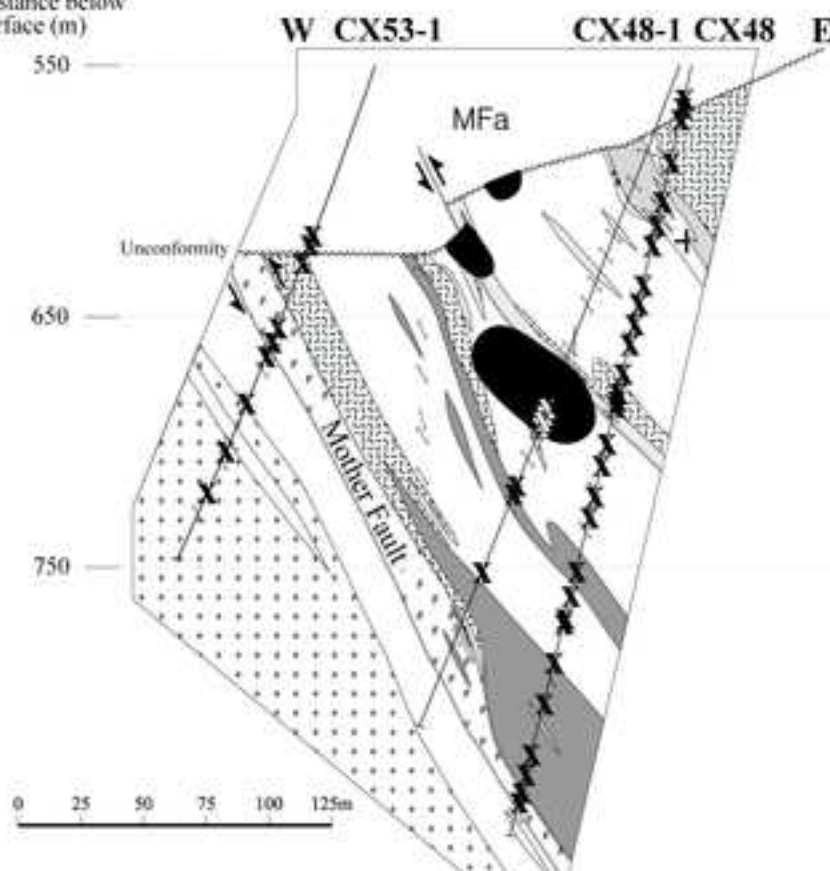
Figure 02



Lithology

A.

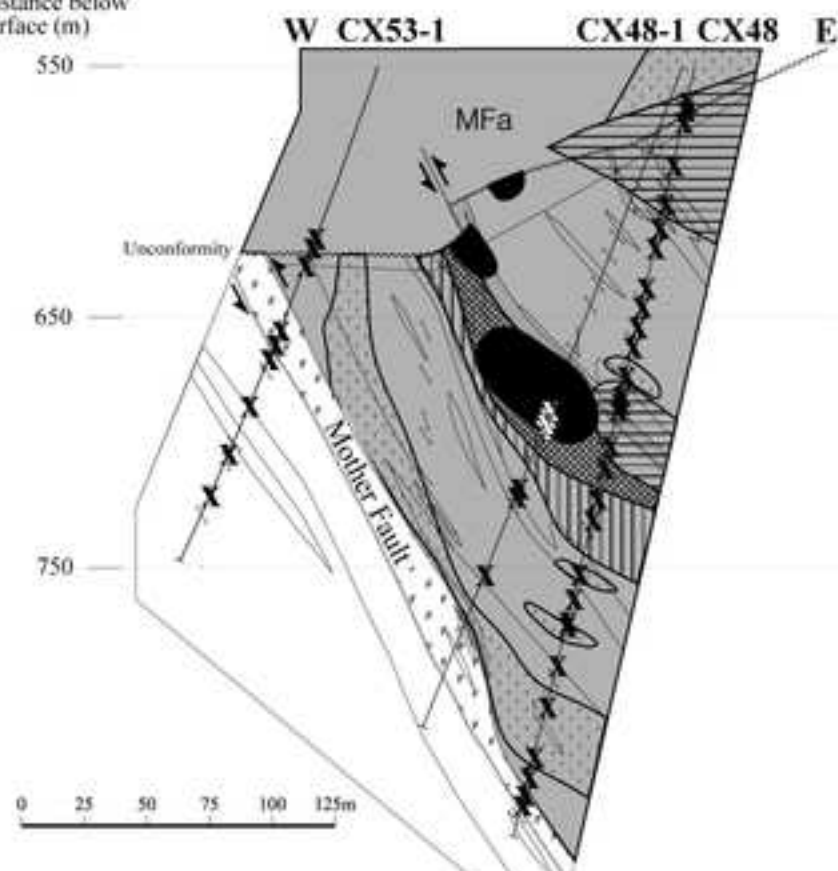
Distance below surface (m)



Alteration

B.

Distance below surface (m)



LEGEND

Mineralization

ATHABASCA GROUP

Manitou Falls Formation

WOLLASTON GROUP

- Pelitic to semipelitic Gneisses
- Wk to Mod Graphitic Pelite/Semipelite
- Cd Gf Pelite (marker)
- Calc-Pelitic/Semipelitic Gneisses
- Pegmatite
- Granite

ALTERATION

- No alteration
- Muscovite alteration
- Chamoisite alteration
- Hematite alteration
- Sudoite alteration
- Dravite alteration

SYMBOLS

- Geological Contacts
- Fault Zone, Fault
- Sample location

Figure 04

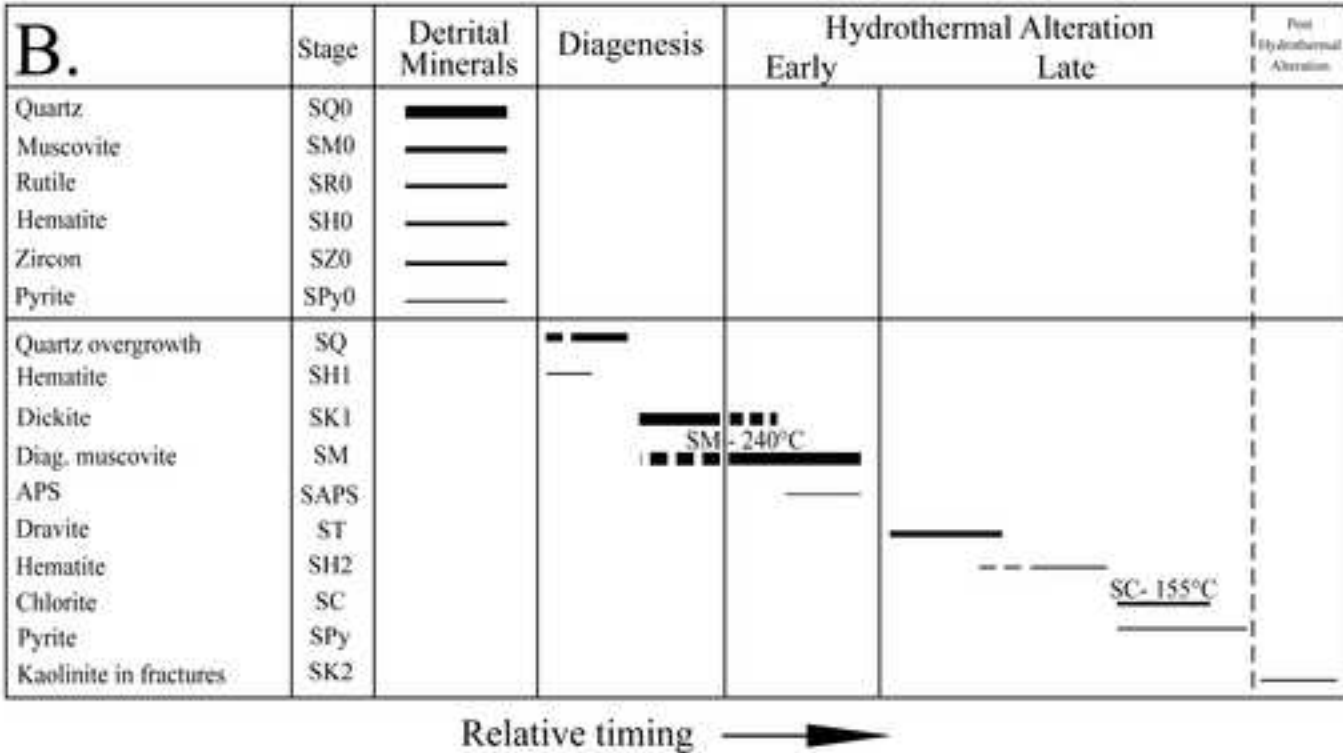
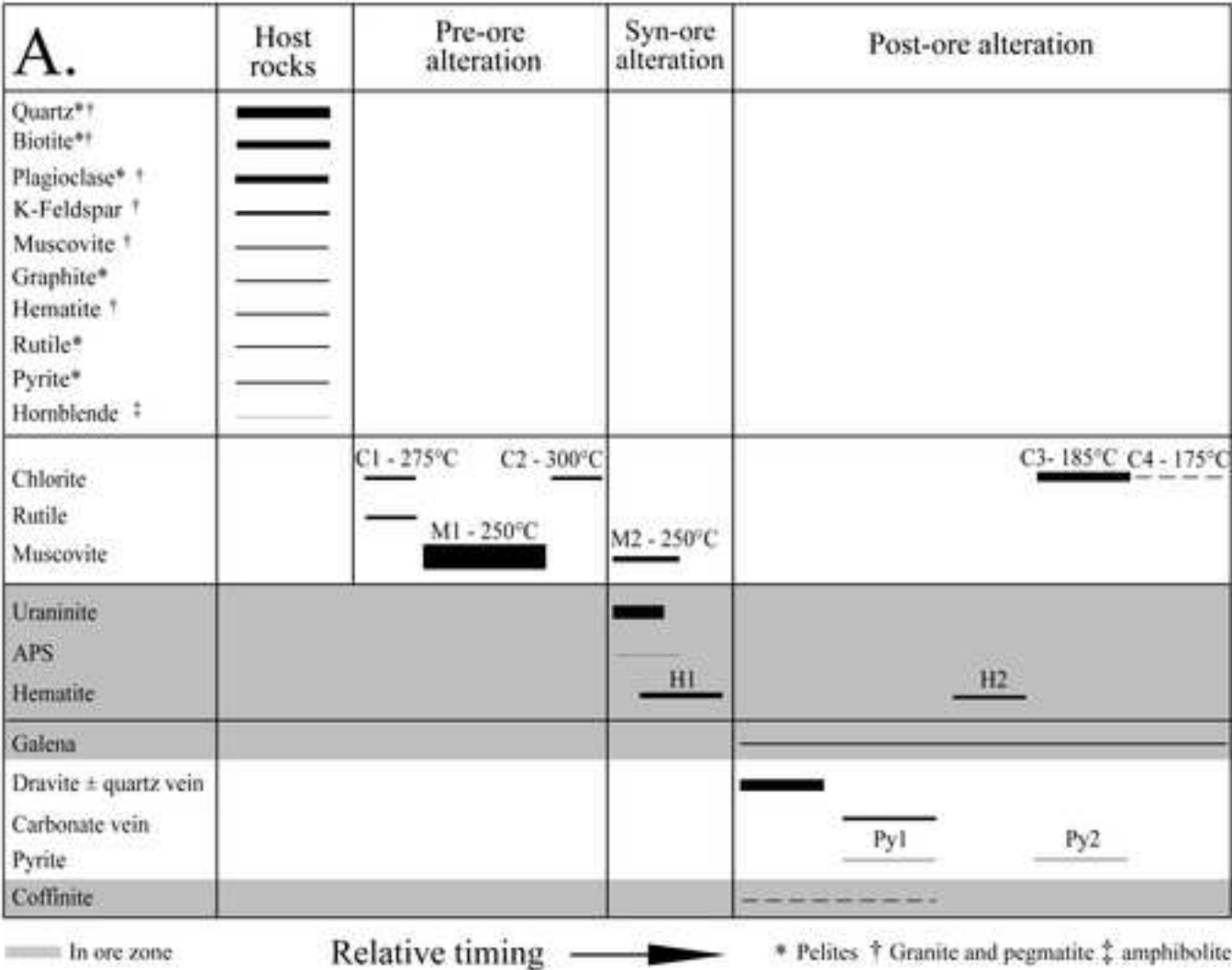


Figure 05

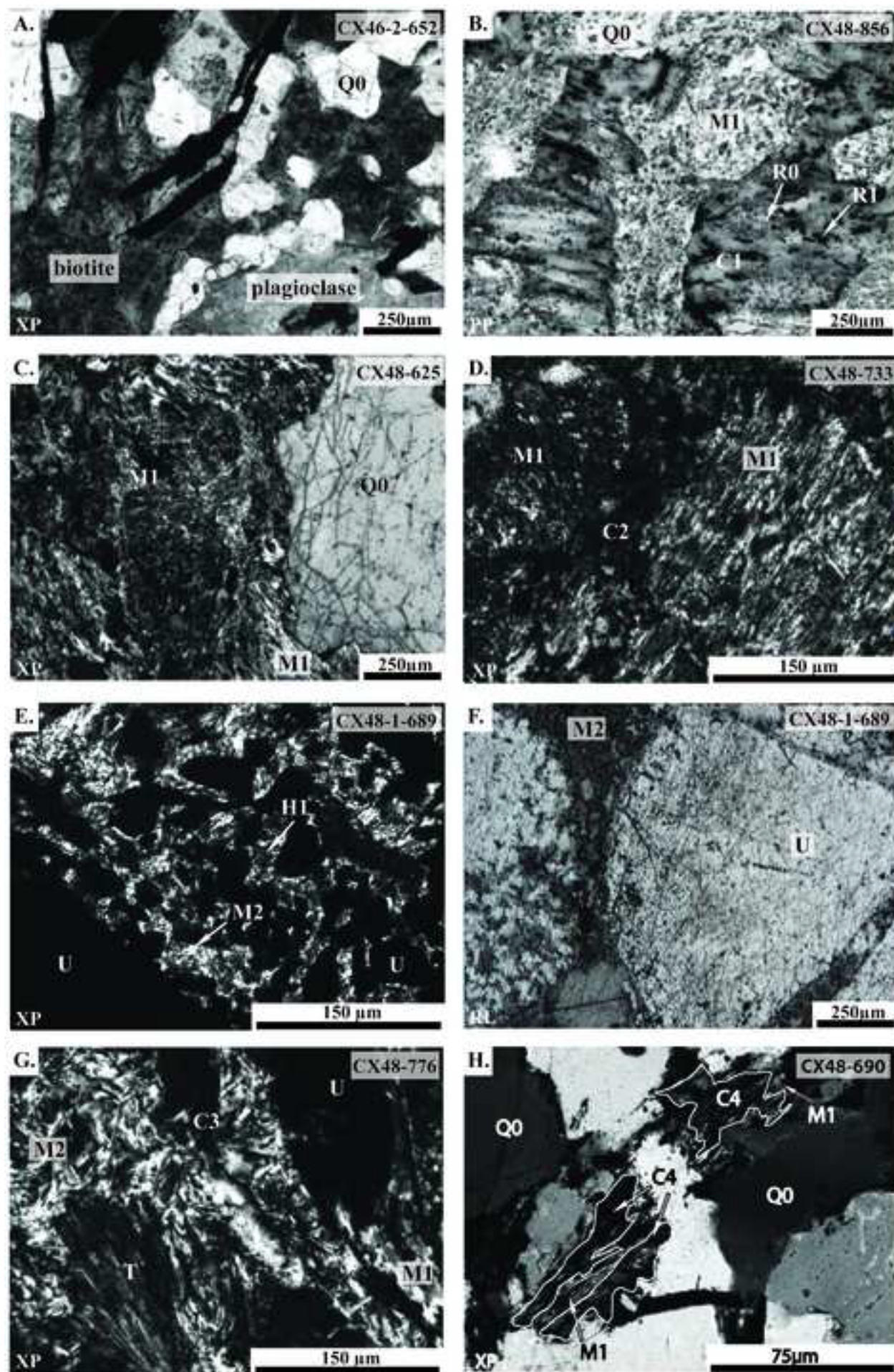


Figure 06

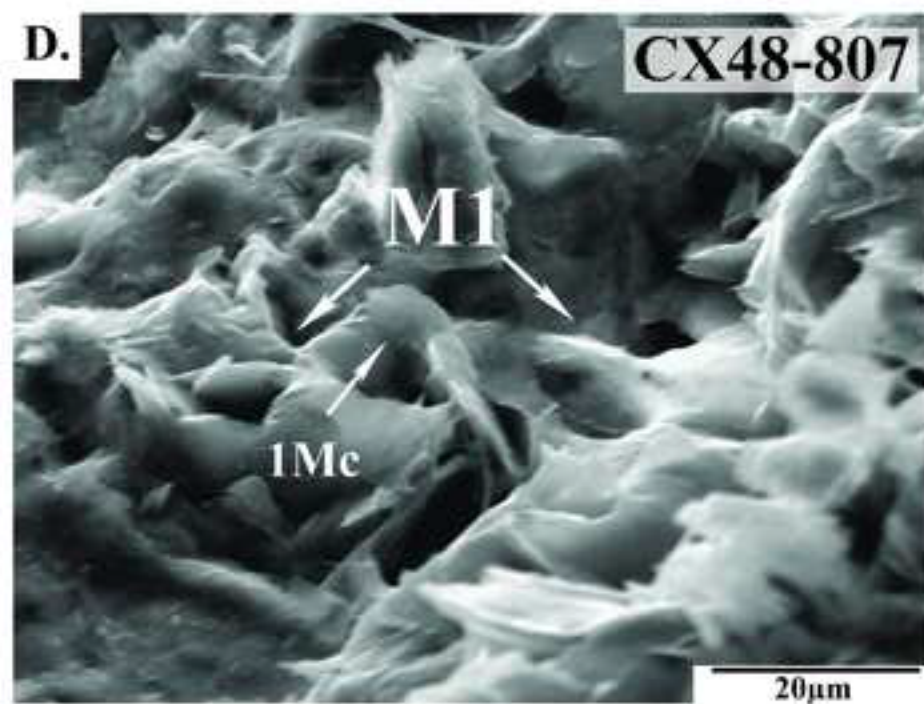
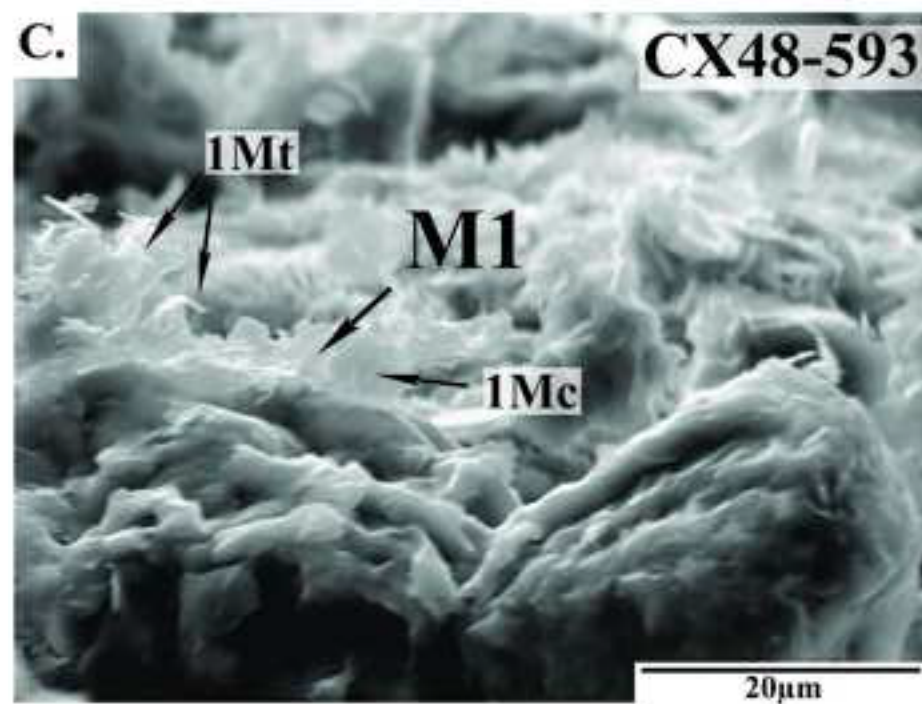
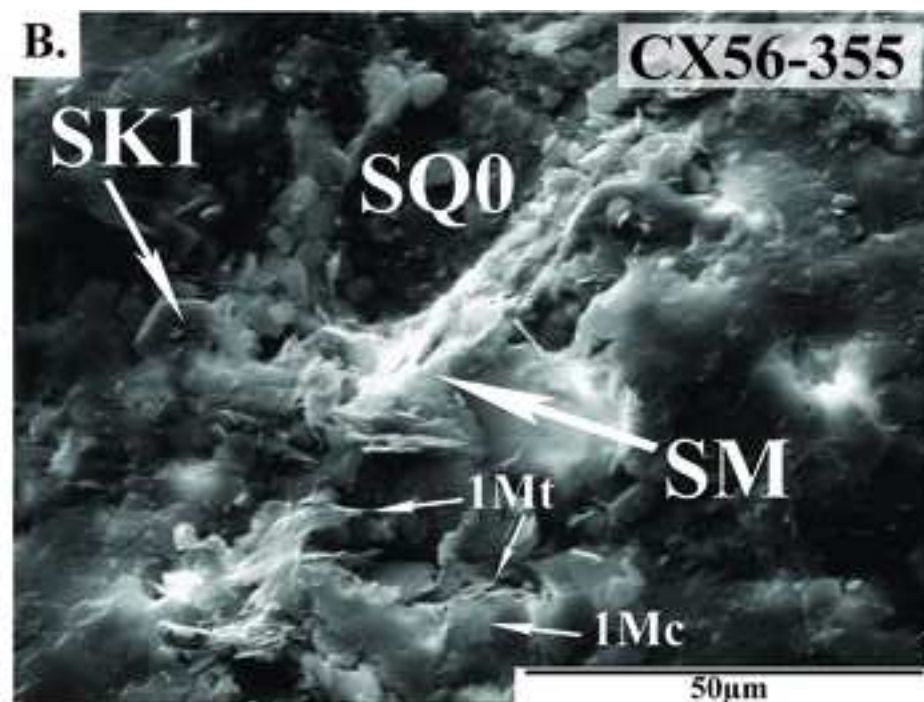
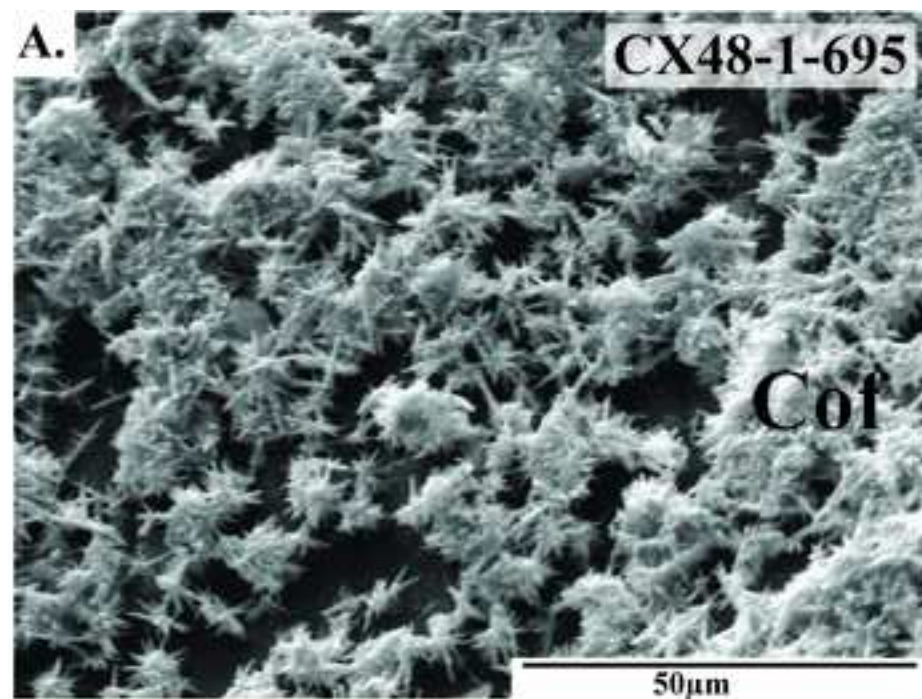


Figure 07

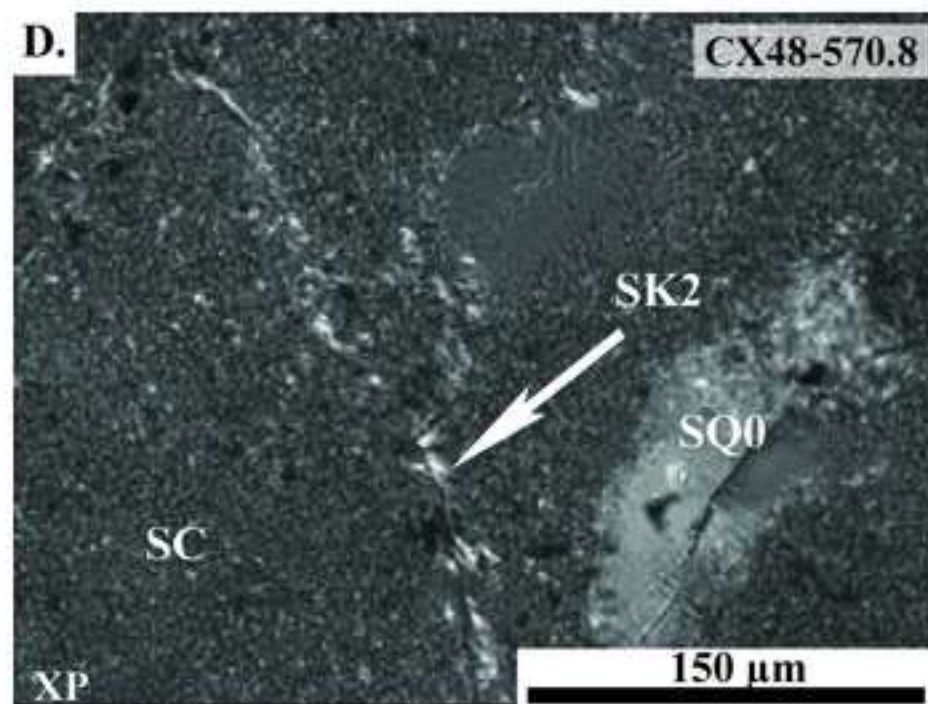
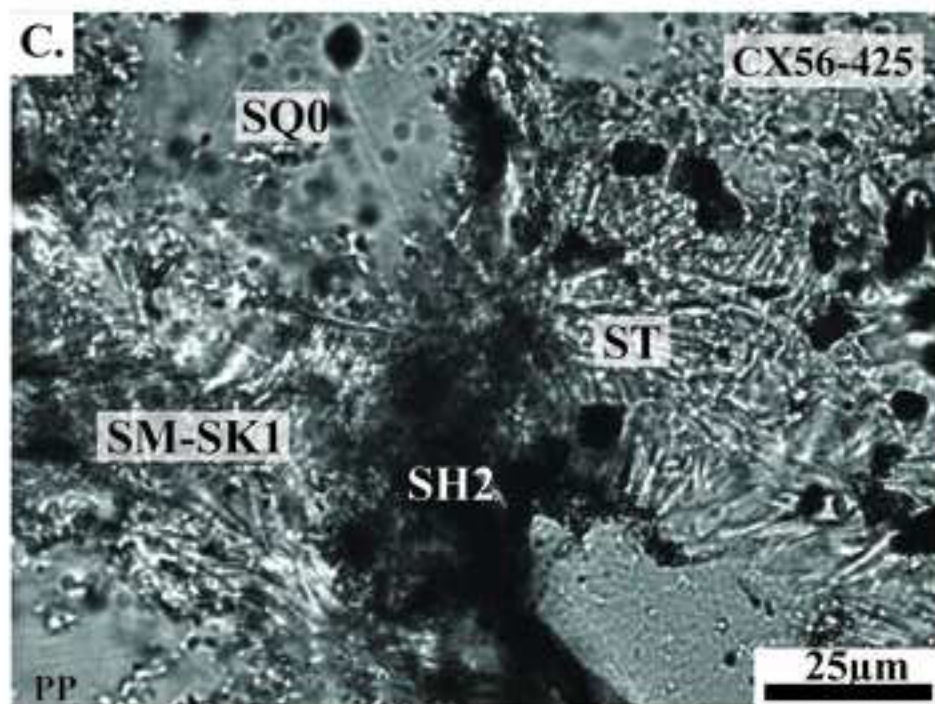
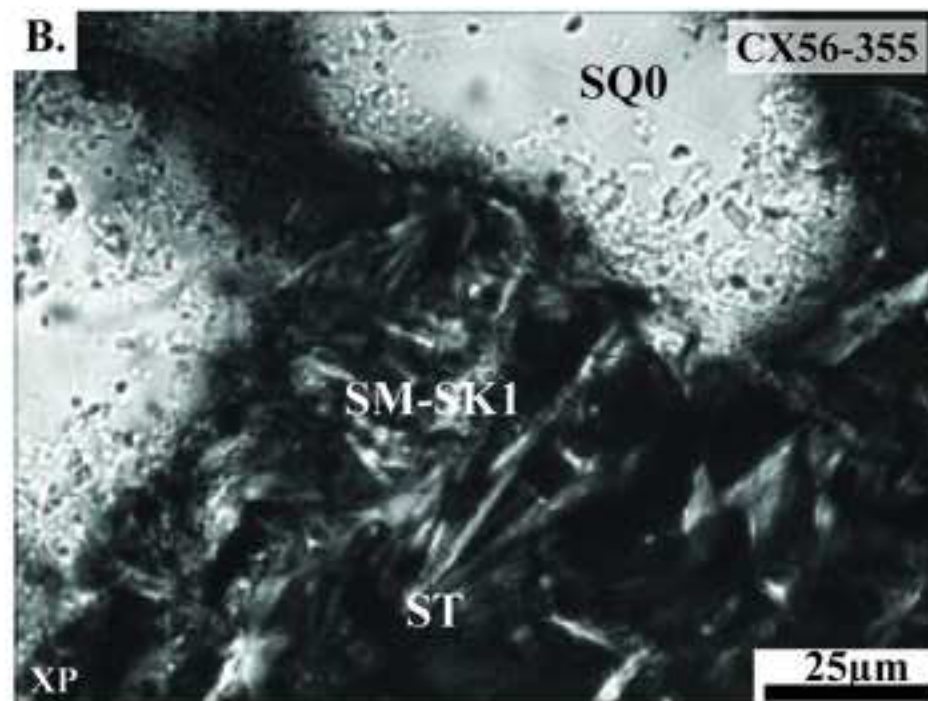
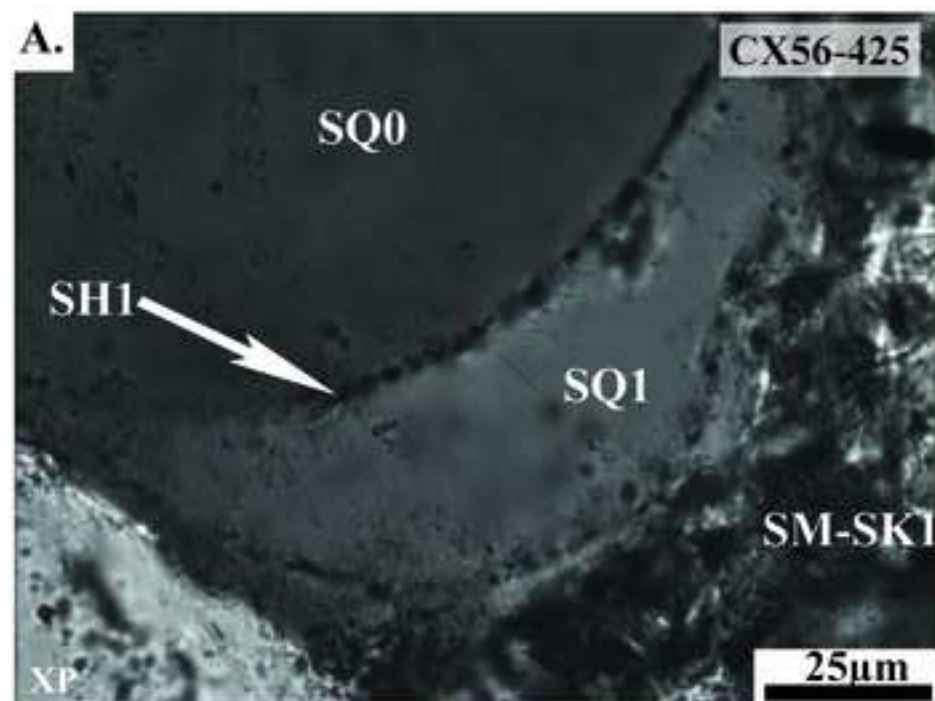


Figure 08

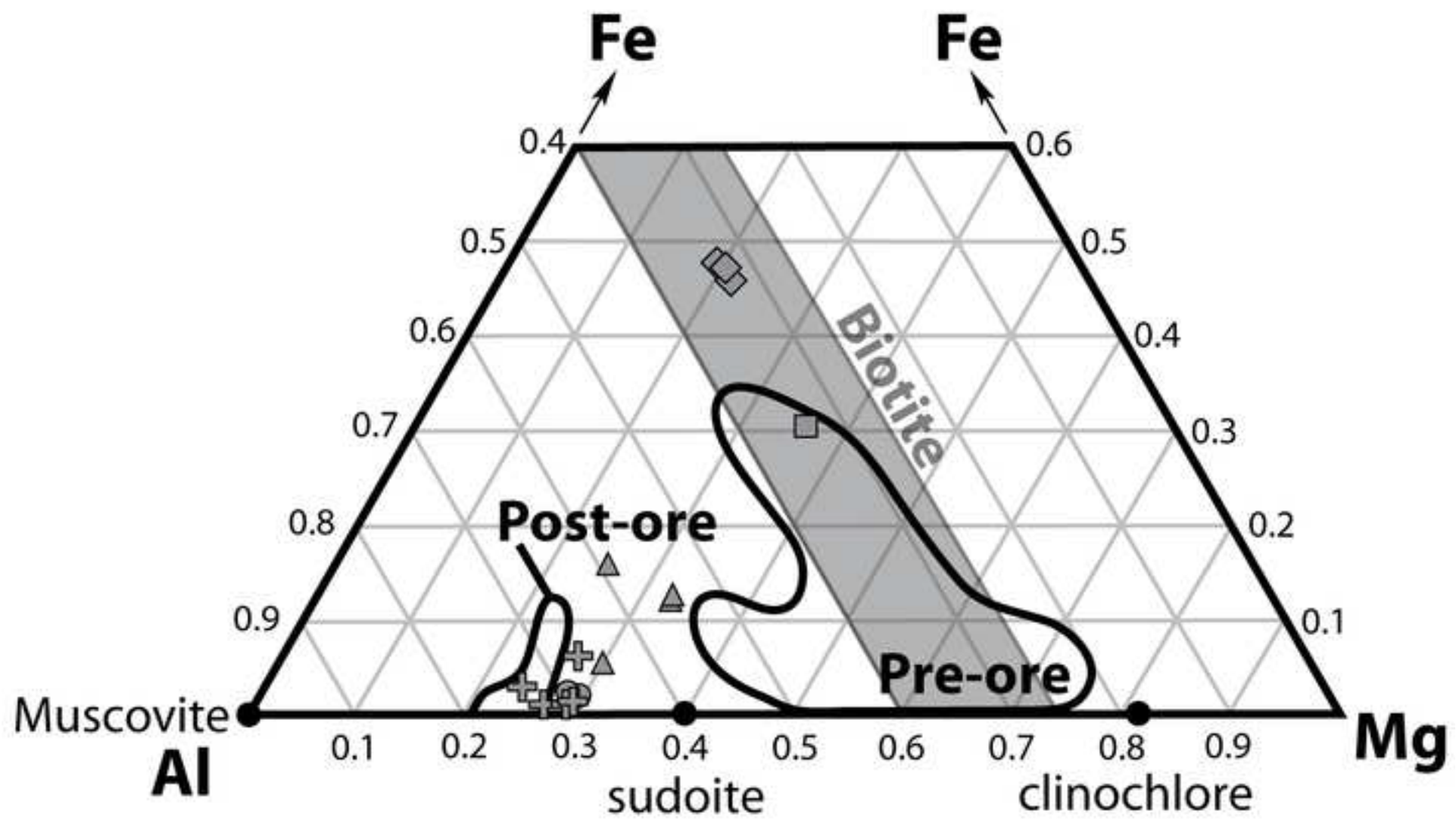


Figure 09

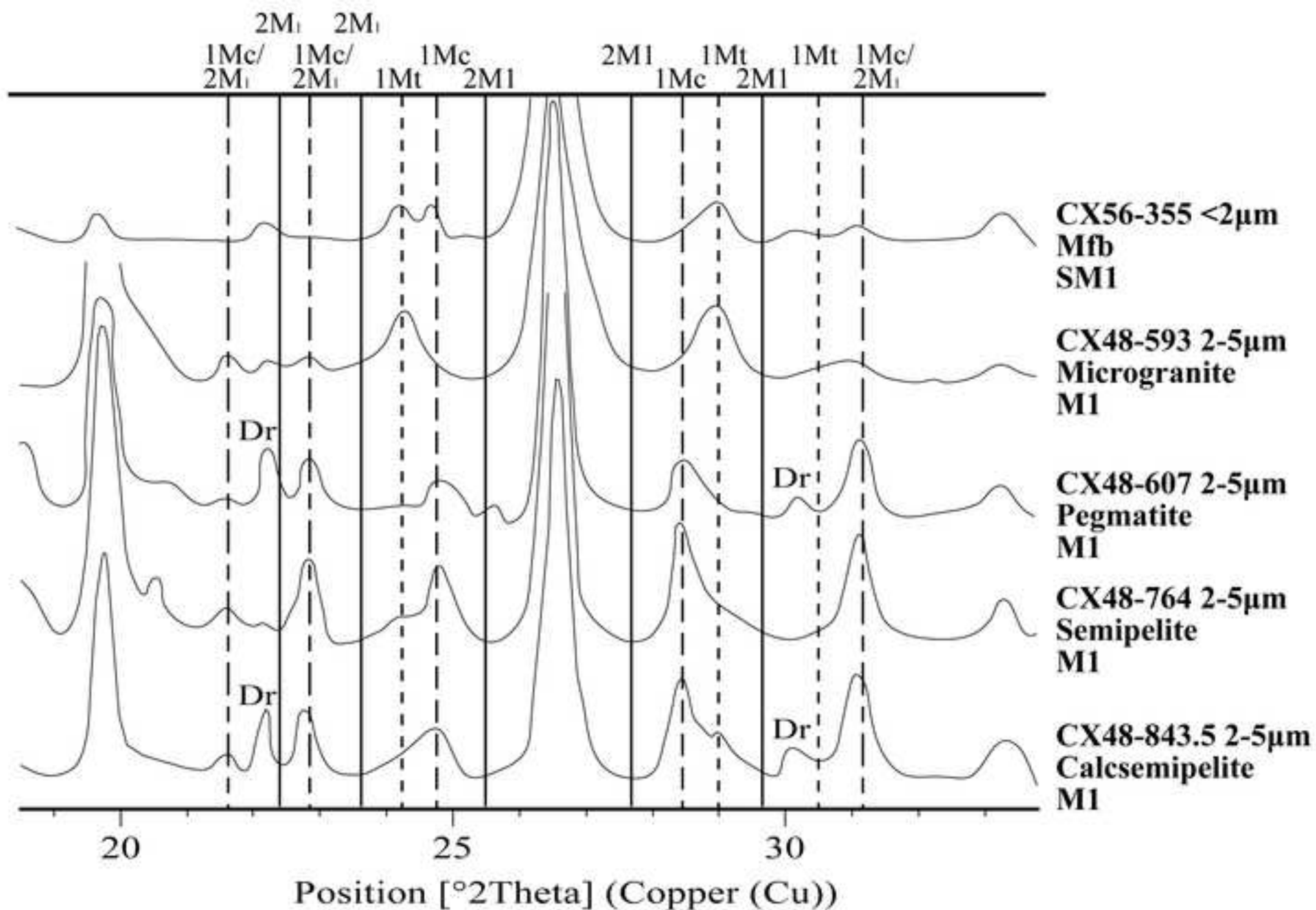


Figure 10

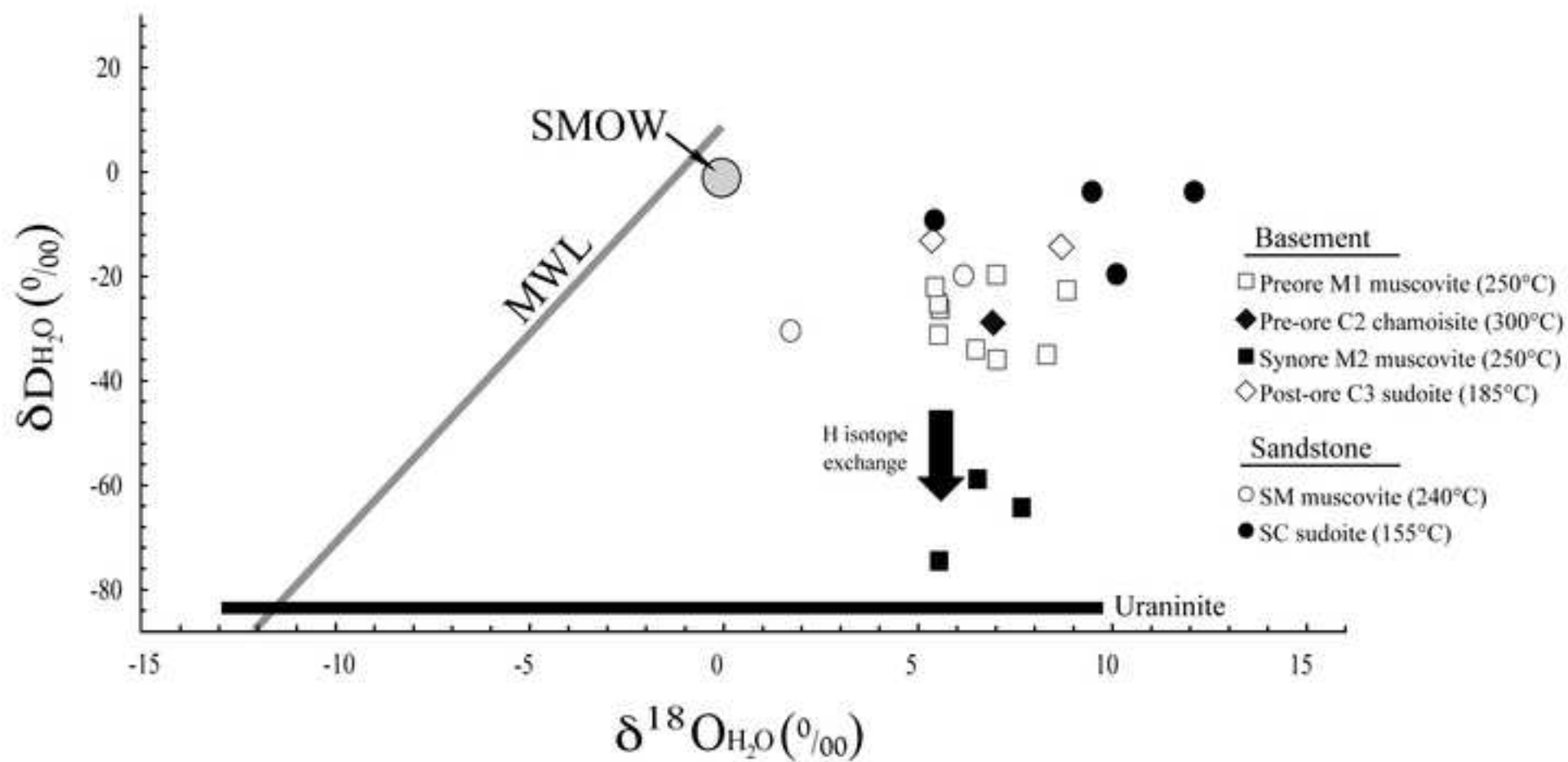


Figure 11

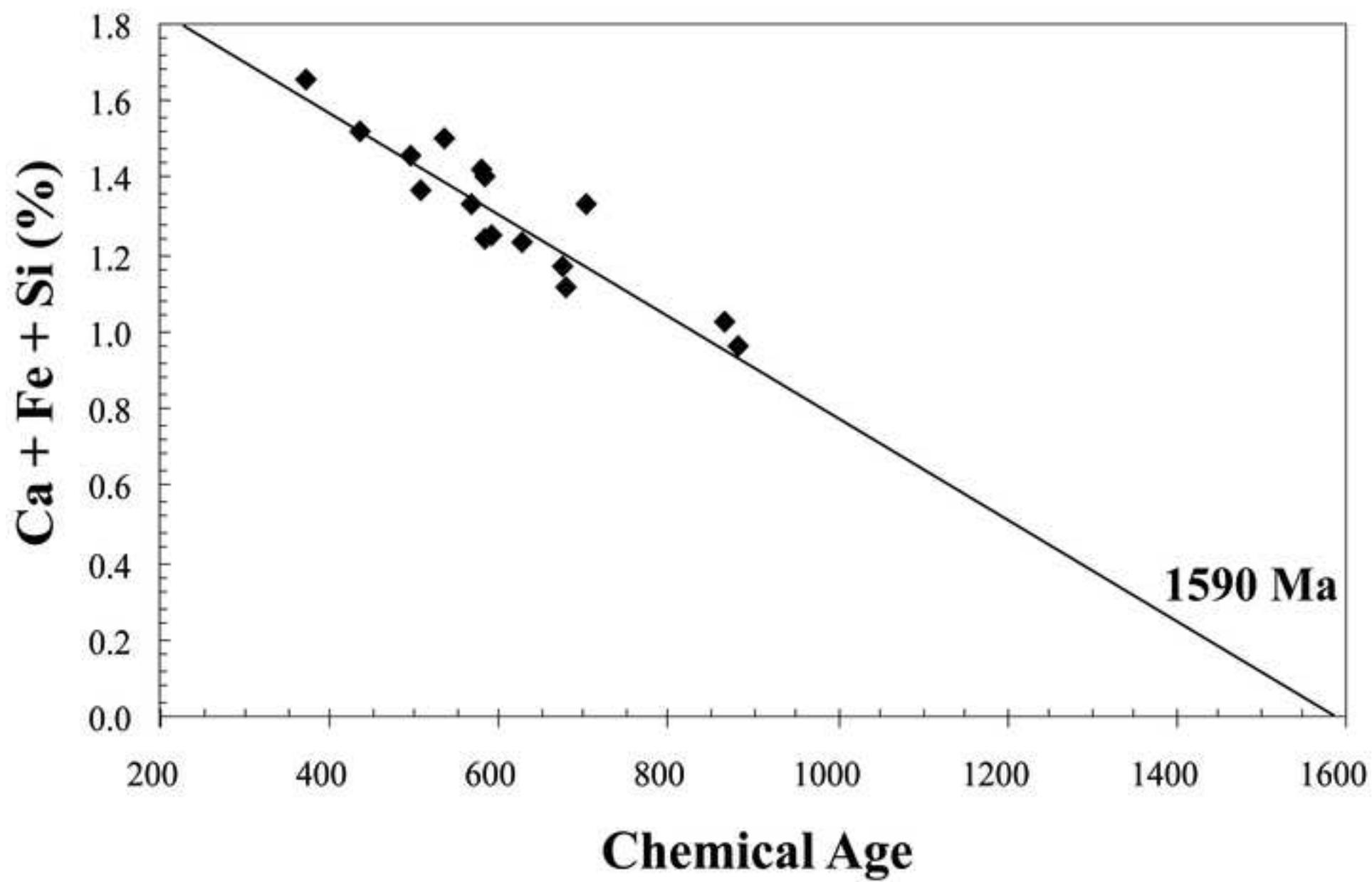


Figure 12

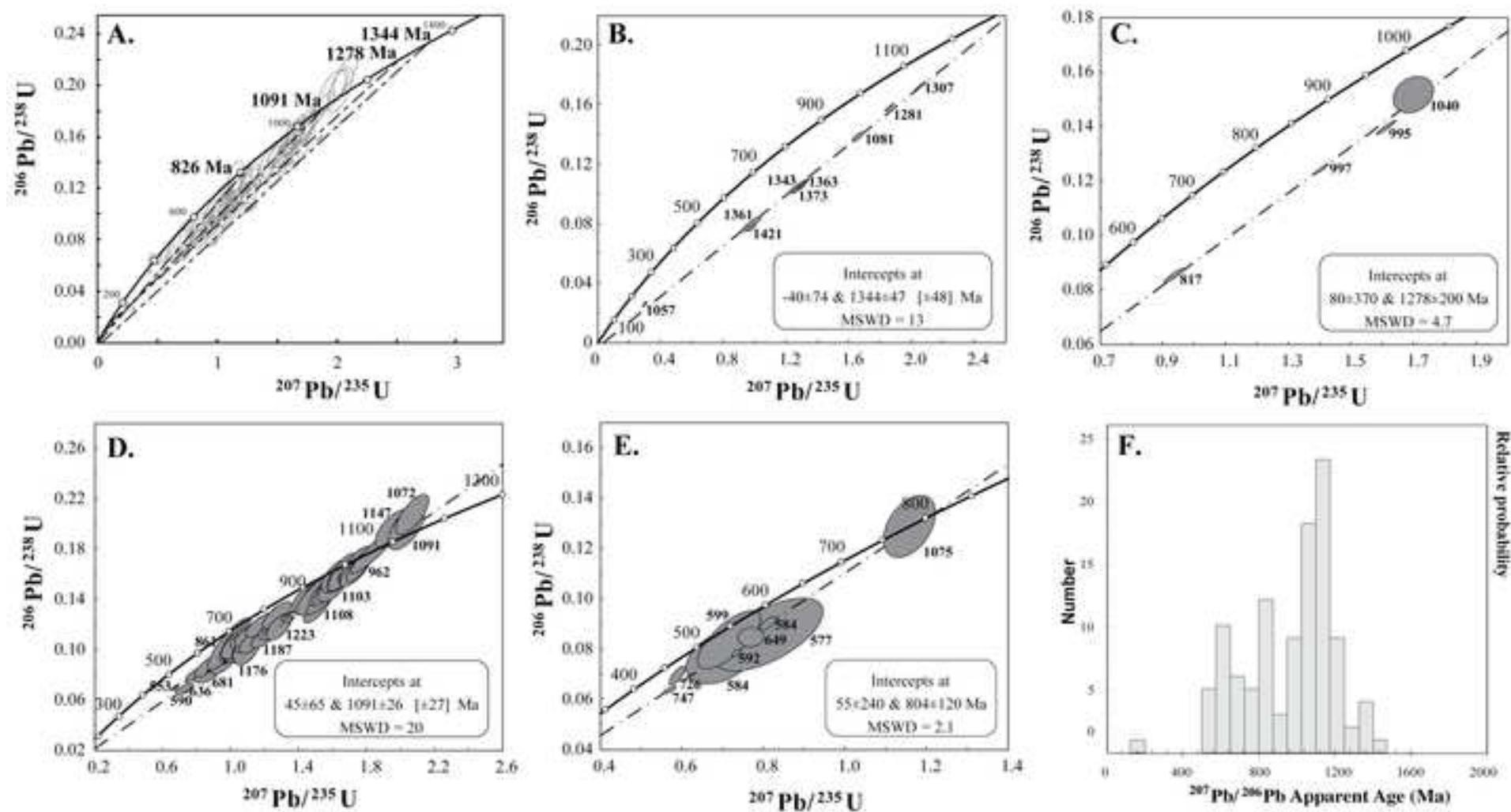


Figure 13

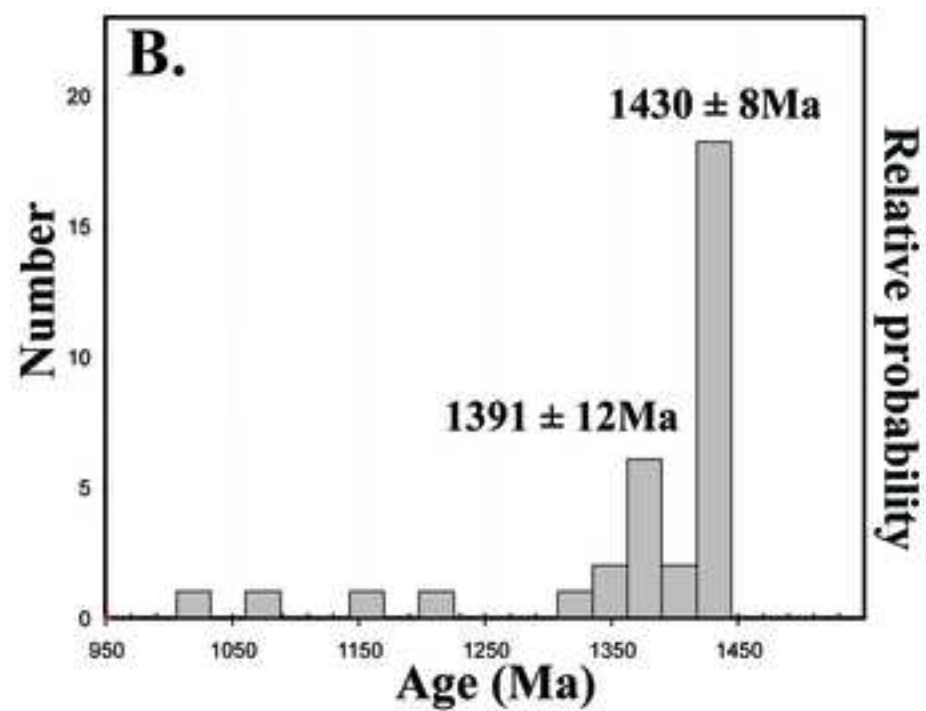
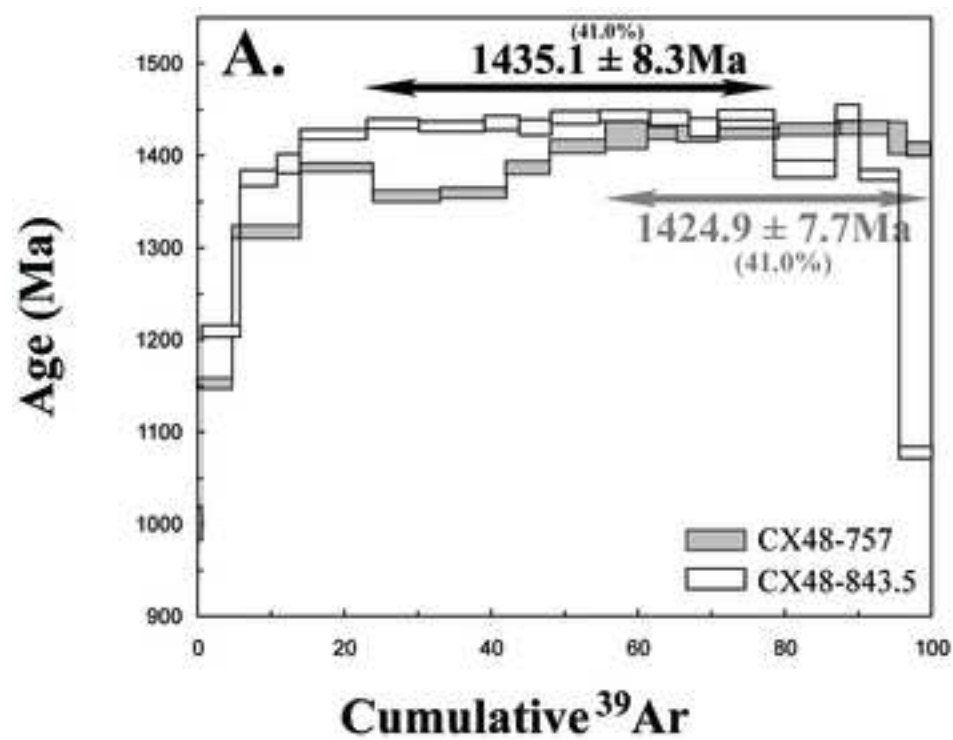
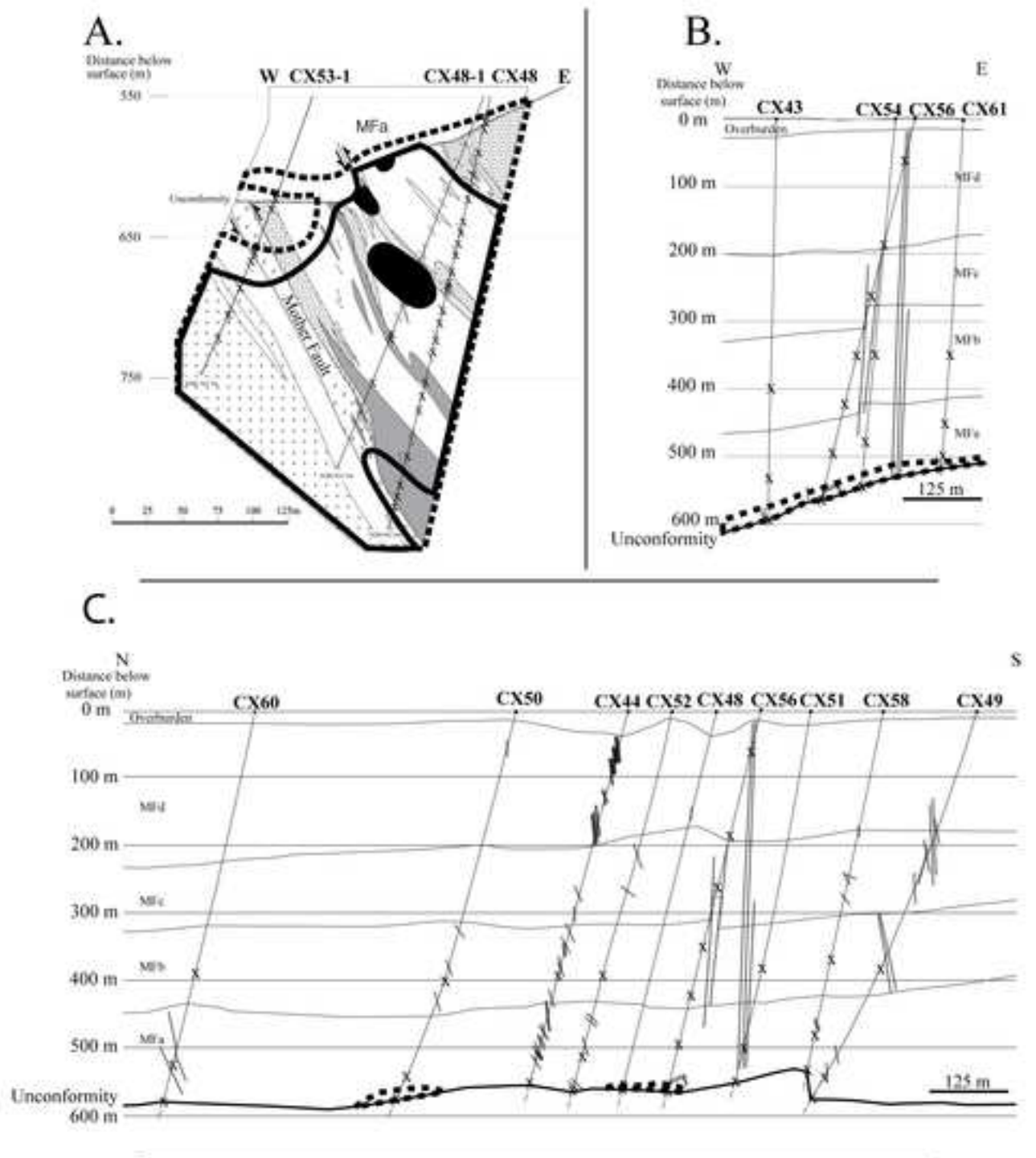


Figure 14



LEGEND

Mineralization

ATHABASCA GROUP

Manitou Falls Formation

WOLLASTON GROUP

- Pelitic to semipelitic Gneisses
- Wk to Mod Graphitic Pelite/Semipelite
- Cd Gf Pelite (marker)
- Calc-Pelitic/Semipelitic Gneisses
- Pegmatite
- Granite

SYMBOLS

- Geological Contacts
- Fault Zone, Fault
- Sample location

$^{206}\text{Pb}/^{204}\text{Pb}$ 30 - 50 $^{206}\text{Pb}/^{204}\text{Pb}$ > 50

Table 01

Table 1: Average Chemical Composition (in wt % and 1σ), Temperatures of Formation, and Average Structural Formulas of Various Muscovite and Chlorite Phases from the Millennium Deposit.

	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	SUM	Temp. (°C)	Average half structural formula
Early hydrothermal alteration muscovite (SM)													
(n=8)	47.5 ± 1.3	<DL	32.2 ± 0.7	<DL	0.7 ± 0.1	<DL	1.2 ± 0.5	0.1 ± 0.1	<DL	9.1 ± 0.5	91.0 ± 1.6	240	K _{0.80} Mg _{0.12} Fe _{0.04} Al _{1.86} (Si _{3.26} ,Al _{0.74})O ₁₀ (OH) ₂
Late hydrothermal alteration sudoite (SC)													
(n=5)	38.4 ± 1.3	<DL	35.4 ± 0.6	<DL	0.6 ± 0.2	<DL	11.7 ± 1.6	0.1 ± 0.1	<DL	-	86.5 ± 2.3	155	Mg _{3.09} Fe _{0.09} Al _{6.23} (Si _{6.82} ,Al _{1.18})O ₂₀ (OH) ₁₆
Pre-ore alteration muscovite (M1)													
(n=9)	50.5 ± 1.3	<DL	33.3 ± 1.7	<DL	0.6 ± 0.5	<DL	1.8 ± 0.8	0.1 ± 0.1	<DL	9.7 ± 0.7	96.2 ± 0.9	250	K _{0.80} Mg _{0.17} Fe _{0.03} Al _{1.82} (Si _{3.27} ,Al _{0.73})O ₁₀ (OH) ₂
Syn-ore alteration muscovite (M2)													
(n=4)	50.1 ± 1.3	<DL	32.7 ± 1.4	<DL	0.9 ± 0.4	<DL	1.7 ± 0.6	0.2 ± 0.1	<DL	8.9 ± 1.2	94.8 ± 0.5	250	K _{0.75} Mg _{0.17} Fe _{0.03} Al _{1.83} (Si _{3.29} ,Al _{0.71})O ₁₀ (OH) ₂
Pre-ore alteration clinocllore (C1)													
(n=1)	28.0	<DL	18.5	<DL	23.7	<DL	15.7	<DL	<DL	<DL	86.3	275	Mg _{5.92} Fe _{4.20} Al _{2.54} (Si _{5.92} ,Al _{2.08})O ₂₀ (OH) ₁₆
Pre-ore alteration chamoisite (C2)													
(n=3)	25.6 ± 0.7	<DL	17.5 ± 0.4	<DL	35.5 ± 0.9	0.5 ± 0.1	8.3 ± 0.6	<DL	<DL	<DL	87.5 ± 0.8	300	Mg _{2.78} Fe _{6.65} Al _{2.36} (Si _{5.73} ,Al _{2.27})O ₂₀ (OH) ₁₆
Post-ore alteration sudoite (C3)													
(n=4)	35.8 ± 2.1	<DL	29.6 ± 2.7	<DL	7.4 ± 3.8	<DL	13.8 ± 1.2	0.4 ± 0.4	<DL	0.4 ± 0.4	87.5 ± 0.8	185	K _{0.08} Mg _{3.45} Fe _{1.16} Al _{5.06} (Si _{6.61} ,Al _{1.39})O ₂₀ (OH) ₁₆
Post-ore alteration sudoite (C4)													
(n=2)	36.1 ± 1.1	<DL	36.4 ± 0.6	<DL	1.6 ± 0.3	<DL	11.9 ± 0.4	<DL	<DL	<DL	86.3 ± 1.6	175	Mg _{2.80} Fe _{0.22} Al _{6.19} (Si _{6.48} ,Al _{1.52})O ₂₀ (OH) ₁₆

Notes: Paragenesis of all phases analysed is shown in Figure 4; the variation of calculated temperatures is ± 30°C; see text for more discussion; n indicates the number of individual analyses on which the average was calculated; <DL = concentration lower than the detection limit; Late drothermal alteration sudoite (SC) were corrected for their muscovite content.

Table 2: Chemical Composition (in wt %) of Aluminum Phosphate-Sulfate (APS) Phases from the Millennium Deposit.

	F	BaO	CaO	FeO	SrO	PbO ₂	ThO ₂	UO ₂	SO ₃	Al ₂ O ₃	La ₂ O ₃	Ce ₂ O ₃	Pr ₂ O ₃	Nd ₂ O ₃	V ₂ O ₃	P ₂ O ₅	Total
	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)	(%)
Early hydrothermal alteration APS (SAPS)																	
CX53-1-664_01	0.53	1.09	1.98	0.16	10.59	0.26	0.23	<DL	4.70	30.46	2.70	4.37	0.30	1.28	<DL	23.31	82.27
CX53-1-664_02	0.71	0.63	2.02	4.27	9.64	<DL	0.03	<DL	10.29	29.22	2.87	4.96	0.38	1.45	0.02	22.44	89.33
CX53-1-664_03	0.46	0.36	2.18	0.15	9.85	<DL	0.09	<DL	5.50	33.33	3.70	5.62	0.28	1.33	0.02	24.72	87.95
CX53-1-664_04	0.44	0.41	2.20	0.22	9.50	0.02	0.05	0.13	5.05	31.41	3.51	5.68	0.52	1.56	0.05	23.05	85.10
CX53-1-664_05	0.56	0.56	1.95	0.11	10.08	<DL	0.03	0.05	5.77	33.33	3.84	5.72	0.45	1.26	<DL	23.30	87.54
CX53-1-664_06	0.71	1.00	1.95	0.15	10.66	<DL	0.11	<DL	5.50	34.56	3.35	5.17	0.30	1.40	0.04	23.39	88.72
CX53-1-664_07	0.86	0.80	2.07	0.28	9.96	0.02	0.09	<DL	5.17	34.42	3.41	6.02	0.37	1.16	<DL	24.00	88.91
Syn-ore APS (APS)																	
CX53-1-622.5_01	0.94	0.73	2.06	5.01	6.11	<DL	0.53	<DL	6.46	29.92	4.70	6.38	0.42	1.01	<DL	22.94	87.61
CX53-1-622.5_02	0.84	0.57	2.22	2.72	8.25	<DL	0.85	<DL	4.31	32.90	3.51	6.53	0.41	1.98	<DL	22.36	87.87
CX53-1-622.5_03	0.92	0.86	2.16	2.82	7.32	<DL	0.89	<DL	4.14	30.41	4.00	6.05	0.45	1.14	0.03	23.14	84.72
CX53-1-622.5_05	0.90	0.82	2.18	2.85	7.08	<DL	0.68	0.07	3.93	30.34	4.39	6.18	0.62	1.05	0.01	22.80	84.49
CX53-1-622.5_04	1.11	0.78	2.32	2.70	7.30	0.01	0.79	0.02	4.30	30.47	3.70	6.19	0.26	1.17	0.03	21.37	83.03
CX53-1-622.5_06	0.72	0.55	1.83	1.40	9.25	0.03	0.10	0.03	4.53	28.47	2.66	5.17	0.51	1.33	0.02	21.53	79.68
CX53-1-622.5_07	0.52	0.37	1.82	0.64	9.04	<DL	0.27	<DL	4.31	26.90	2.38	4.63	0.42	1.69	<DL	20.66	77.29
CX53-1-617.5_01	0.64	0.70	1.91	1.13	10.84	0.02	0.45	<DL	4.64	30.73	1.80	4.27	0.20	1.20	0.01	21.36	81.69
CX53-1-617.5_02	0.75	0.65	1.97	1.62	9.99	<DL	0.32	<DL	4.92	28.96	1.69	4.04	0.47	1.06	0.04	19.52	79.18
CX53-1-617.5_03	0.60	0.50	1.68	1.10	6.57	<DL	0.38	0.14	3.27	25.62	1.57	2.90	0.18	0.92	0.04	13.92	69.12
CX53-1-617.5_04	0.74	0.75	2.29	1.31	10.39	<DL	0.70	<DL	5.27	30.72	2.60	4.05	0.14	1.19	0.01	21.91	83.29
CX53-1-617.5_05	0.91	0.72	2.01	1.10	10.71	0.04	0.95	0.01	5.42	30.47	1.74	4.33	0.39	1.59	0.01	21.92	82.74
CX53-1-617.5_06	0.64	0.59	1.96	0.91	9.51	0.01	0.68	0.04	5.41	27.58	1.77	3.76	0.35	1.61	0.03	19.61	76.86
CX53-1-617.5_07	0.61	0.69	1.55	0.69	10.05	<DL	0.23	0.20	4.34	25.98	1.85	3.78	0.29	1.61	<DL	20.59	72.92
CX53-1-617.5_08	0.62	0.63	1.61	0.65	10.31	0.01	0.14	0.05	4.62	28.06	1.90	3.89	0.26	1.81	<DL	21.08	76.05
CX56-425_01	0.67	0.64	1.97	0.51	10.52	<DL	0.30	<DL	4.61	31.98	2.13	4.48	0.50	1.43	<DL	23.23	83.58
CX56-425_02	0.95	0.99	2.77	0.91	9.28	0.01	0.38	<DL	3.69	31.17	2.19	4.95	0.25	1.22	<DL	24.02	84.35

Notes: Paragenesis of all phases analysed is shown in Figure 4; see text for more discussion; <DL = concentration lower than the detection limit. APS compositions have been corrected for their muscovite content.

Table 3. Uraninite Average Chemical Composition and Calculated Chemical Age from the Millennium Deposit.

	U	Pb	Si	Fe	Mg	Ca	Ti	V	O	Total
CX48-1-684.5-3	69.8	2.3	1.5	<DL	<DL	0.3	3.8	0.4	14.3	92.4
CX48-1-684.5-5	71.1	4.0	0.9	0.2	<DL	0.4	2.3	0.2	13.0	92.2
CX48-1-684.5-7	74.7	6.6	<DL	0.2	<DL	1.3	0.0	<DL	11.8	94.9
CX48-1-684.5-8	72.5	8.8	<DL	0.2	<DL	1.1	0.1	<DL	11.8	94.8
CX48-1-684.5-9	75.1	4.7	<DL	<DL	<DL	1.4	0.0	0.2	11.7	93.4
CX48-1-689-1	72.9	7.3	<DL	0.2	<DL	1.1	0.4	<DL	11.9	94.0
CX48-1-689-2	74.1	6.2	<DL	<DL	<DL	1.2	0.3	<DL	11.8	94.0
CX48-1-689-3	75.4	5.5	<DL	<DL	<DL	1.4	0.2	0.2	11.9	94.6
CX48-1-689-4	71.2	10.8	<DL	<DL	<DL	0.8	0.1	<DL	11.8	95.1
CX48-1-689-5	71.1	11.1	<DL	<DL	<DL	0.8	0.1	<DL	11.8	95.2
CX48-1-689-6	74.4	6.4	<DL	<DL	<DL	1.2	0.1	<DL	11.7	94.2
CX48-1-689-7	76.4	0.8	1.8	0.3	<DL	0.9	0.1	0.3	13.2	93.8
CX48-1-689-8	73.9	7.5	<DL	<DL	<DL	1.1	0.1	<DL	11.8	94.6
CX48-1-689-9	73.8	7.1	<DL	<DL	<DL	1.2	0.2	<DL	11.8	94.4
CX48-1-689-10	73.7	7.3	<DL	<DL	<DL	1.1	0.2	<DL	11.8	94.3
CX48-1-689-11	73.4	8.6	<DL	<DL	<DL	1.0	0.2	<DL	11.8	95.1
CX48-1-689-12	74.3	6.7	<DL	<DL	<DL	1.2	0.2	<DL	11.9	94.7
CX48-1-689-13	73.2	8.5	<DL	<DL	<DL	1.0	0.1	<DL	11.8	95.0
CX48-1-689-14	74.5	8.0	<DL	<DL	<DL	1.1	0.2	<DL	12.0	96.1
CX48-1-689-15	74.6	7.4	<DL	<DL	<DL	1.2	0.3	<DL	12.1	95.9
average:	73.3	6.4	0.4	0.1	<DL	1.0	0.6	0.2	12.2	94.1
±1σ	1.9	2.9	0.6	0.1		0.4	1.1	0.1	0.8	0.9

Table 4: Measured Mineral $\delta^{18}\text{O}$ and δD and Calculated $\delta^{18}\text{O}$ and δD Values for Fluids in Equilibrium with Alteration Minerals and Uraninite from the Millennium Deposits.

Sample	fraction size	Mineral		Water	
		δD	$\delta^{18}\text{O}$	δD	$\delta^{18}\text{O}$
Early hydrothermal alteration muscovite (SM)					
CX49-385	<2	-50	11.3	-20	6.1
CX56-499.5	<2	-61	7.0	-31	1.8
Late hydrothermal alteration sudoite (SC)					
CX48-567,1	<2	-53	9.3	-9	5.5
CX54-598	<2	-48	15.9	-4	12.1
CX56-186.5	<2	-64	13.9	-20	10.1
CX60-596	<2	-48	13.4	-4	9.6
Pre-ore muscovite (M1)					
CX48-593	<2	-65	13.1	-35	8.3
CX48-607	2-5	-50	11.8	-20	7.0
CX48-625	<2	-66	11.8	-36	7.0
CX48-688	<2	-53	13.7	-23	8.9
CX48-704.5	<2	-55	10.2	-25	5.4
CX48-764	2-5	-61	10.3	-31	5.5
CX48-807.5	2-5	-52	10.1	-22	5.3
CX53-1-629	<2	-56	10.3	-26	5.5
Pre-ore chamoisite (C2)					
CX48-733	<2	-54	7.1	-29	7.0
Syn-ore muscovite (M2)					
CX48-1-684	-	-89	11.4	-59	6.6
CX48-1-689	-	-94	12.6	-64	7.8
CX48-1-695	-	-104	10.2	-74	5.4
Uraninite					
CX48-1-689		-12.9			
CX48-1-695		-8.6			
CX56-3-680.5		9.8			
CX56-3-711		-0.8			
Post-ore sudoite (C3)					
CX48-713	<2	-55	8.1	-13	5.4
CX53-1-686	<2	-57	11.5	-15	8.8

Notes: The temperatures used to calculate the fluid values are derived from the crystal chemistry of the clay minerals; the variation of the individual O and H analyses are ± 0.2 and ± 3 per mil; see the text for more details

Table 5. Isotopic data and apparent ages for uraninite from the Millennium deposit.

Analytical method	Sample	²⁰⁶ Pb/ ²⁰⁴ Pb	com. Pb* (%)	²⁰⁷ Pb/ ²³⁵ U	±1σ	²⁰⁶ Pb/ ²³⁸ U†	±1σ	R‡	²⁰⁷ Pb/ ²⁰⁶ Pb	±1σ	Apparent ages (±1σ, Ma)**						Disc.†† (%)
											²⁰⁶ Pb/ ²³⁸ U	error	²⁰⁷ Pb/ ²³⁵ U	error	²⁰⁷ Pb/ ²⁰⁶ Pb	error	
<i>Theoretical ratio for pure non disturbed uraninite with an age of 1590 Ma</i>		²⁰⁴ Pb <DL	0.00	3.7870	-	0.2798	-	-	0.0982	-	1590	-	1590	-	1590	-	0
HR-ICP-MS	CX48-1-684_01	67878	0.01	1.4672	0.0687	0.1402	0.0083	0.479	0.0778	0.0028	656	428	747	8	1119	141	41
HR-ICP-MS	CX48-1-684_07	²⁰⁴ Pb <DL	0.00	1.1149	0.0602	0.1103	0.0060	0.511	0.0805	0.0036	696	377	758	6	1035	176	33
HR-ICP-MS	CX48-1-684_A02	118241	0.02	1.4695	0.0102	0.1374	0.0016	0.519	0.0805	0.0005	478	66	534	2	823	26	42
HR-ICP-MS	CX48-1-684_A03	303161	0.01	1.0903	0.0174	0.1073	0.0019	0.910	0.0756	0.0004	788	111	902	2	1152	22	32
HR-ICP-MS	CX48-1-684_A05	123807	0.02	0.5887	0.0130	0.0696	0.0025	0.7645	0.0636	0.0008	433	83	469	3	726	50	40
HR-ICP-MS	CX48-1-684_A06	97959	0.02	1.3114	0.0096	0.1257	0.0014	0.493	0.0780	0.0005	639	62	716	1	1021	27	37
HR-ICP-MS	CX48-1-684_A07	115748	0.02	1.1137	0.0239	0.1141	0.0026	0.090	0.0738	0.0004	557	17	602	24	827	23	33
HR-ICP-MS	CX48-1-684_A08	91641	0.02	1.2501	0.0139	0.1207	0.0025	0.061	0.0770	0.0009	722	89	835	3	1108	49	35
HR-ICP-MS	CX48-1-684_E01	161921	0.01	1.5423	0.0077	0.1502	0.0013	0.415	0.0760	0.0004	636	49	736	1	1149	19	45
HR-ICP-MS	CX48-1-684_E02	87044	0.02	1.0774	0.0139	0.1077	0.0014	0.005	0.0753	0.0003	797	89	907	1	1143	14	30
HR-ICP-MS	CX48-1-684_E03	168165	0.02	1.3739	0.0096	0.1304	0.0011	0.780	0.0797	0.0003	637	62	692	1	1054	16	40
HR-ICP-MS	CX48-1-684_E04	149934	0.01	1.2265	0.0089	0.1203	0.0013	0.499	0.0768	0.0005	734	57	823	1	1117	28	34
HR-ICP-MS	CX48-1-684_E06	100897	0.02	1.1891	0.0074	0.1165	0.0009	0.584	0.0767	0.0003	805	48	937	1	1224	18	34
HR-ICP-MS	CX48-1-689_12	410354	0.01	1.1582	0.0420	0.1288	0.0082	0.4334	0.0759	0.0033	780	265	780	8	1075	171	27
HR-ICP-MS	CX48-1-689_13	55613	0.01	1.0414	0.0601	0.1020	0.0081	0.284	0.0774	0.0067	470	376	562	8	644	343	27
HR-ICP-MS	CX48-1-689_14	21872	0.01	0.9902	0.0833	0.1045	0.0136	0.854	0.0747	0.0019	760	516	830	14	861	105	12
HR-ICP-MS	CX48-1-689_15	32337	0.04	1.0940	0.0469	0.1073	0.0064	0.591	0.0777	0.0029	842	296	974	7	1248	150	33
HR-ICP-MS	CX48-1-689_18	16326	0.01	1.2131	0.0336	0.1130	0.0033	0.185	0.0765	0.0032	804	213	935	3	1223	169	34
HR-ICP-MS	CX48-1-689_19	13000	0.11	1.1859	0.0652	0.1205	0.0065	0.746	0.0733	0.0024	732	407	812	7	1108	131	34
HR-ICP-MS	CX48-1-689_20	44322	0.01	1.0312	0.0560	0.1044	0.0084	0.713	0.0738	0.0030	569	52	649	8	762	166	25
HR-ICP-MS	CX48-1-689_21	68610	0.00	1.0929	0.0524	0.1018	0.0088	0.473	0.0797	0.0047	761	329	873	9	1126	233	32
HR-ICP-MS	CX48-1-689_22	44322	0.01	1.0312	0.0560	0.1044	0.0084	0.713	0.0738	0.0030	518	352	545	8	697	166	26
HR-ICP-MS	CX48-1-689_23	121601	0.02	1.1891	0.0460	0.1093	0.0051	0.661	0.0768	0.0023	742	290	856	5	1124	122	34
HR-ICP-MS	CX48-1-689_24	70487	0.00	1.0684	0.0408	0.1039	0.0089	0.740	0.0785	0.0024	827	258	862	9	1176	121	30
HR-ICP-MS	CX48-1-689_27	129519	0.03	1.0848	0.0582	0.1079	0.0050	0.451	0.0760	0.0045	829	365	917	5	1205	238	31
HR-ICP-MS	CX48-1-689_C1_01	351966	0.01	2.0693	0.0646	0.2076	0.0105	0.715	0.0585	0.0014	426	404	553	11	1072	103	60
HR-ICP-MS	CX48-1-689_C1_02	1546093	0.00	1.2391	0.0359	0.1237	0.0074	0.748	0.0600	0.0023	727	227	842	8	1114	165	35
HR-ICP-MS	CX48-1-689_C1_03	4922658	0.00	1.8271	0.0425	0.1754	0.0051	0.765	0.0601	0.0011	553	268	678	5	1152	82	52
HR-ICP-MS	CX48-1-689_C1_04	1088172	0.00	1.8396	0.0884	0.1782	0.0115	0.911	0.0611	0.0009	477	546	592	12	1026	67	54
HR-ICP-MS	CX48-1-689_C1_05	608863	0.00	1.7196	0.0583	0.1644	0.0077	0.729	0.0600	0.0015	479	366	580	8	962	109	50
HR-ICP-MS	CX48-1-689_C1_07	²⁰⁴ Pb <DL	0.00	2.0157	0.0527	0.1996	0.0084	0.676	0.0599	0.0012	500	331	626	9	1091	89	54
HR-ICP-MS	CX48-1-689_C1_08	1363763	0.00	2.0298	0.0692	0.1968	0.0116	0.627	0.0589	0.0014	430	431	554	12	1063	101	60
HR-ICP-MS	CX48-1-689_C2_01	4183215	0.00	1.2720	0.0556	0.1276	0.0072	0.769	0.0599	0.0017	607	349	671	7	956	120	36
HR-ICP-MS	CX48-1-689_C2_02	642495	0.00	1.5461	0.0548	0.1468	0.0074	0.797	0.0630	0.0010	569	344	673	7	1040	65	45
HR-ICP-MS	CX48-1-689_C2_03	1636535	0.00	1.6575	0.0547	0.1619	0.0097	0.744	0.0622	0.0021	439	343	520	10	861	147	49
HR-ICP-MS	CX48-1-689_C2_04	630040	0.00	1.6522	0.0580	0.1601	0.0093	0.730	0.0637	0.0022	575	363	707	9	1121	147	49
HR-ICP-MS	CX48-1-689_C2_05	3090785	0.00	1.2690	0.0412	0.1192	0.0042	0.434	0.0626	0.0024	649	261	723	4	1010	160	36
HR-ICP-MS	CX48-1-689_C2_06	220430	0.01	1.5068	0.0549	0.1476	0.0075	0.747	0.0609	0.0016	646	344	758	8	1131	113	43
HR-ICP-MS	CX48-1-689_C2_07	409330	0.00	1.7347	0.0603	0.1635	0.0084	0.764	0.0627	0.0019	438	378	526	8	891	132	51
HR-ICP-MS	CX48-1-689_C2_08	2535204	0.00	1.7329	0.0619	0.1679	0.0095	0.777	0.0607	0.0015	555	387	688	10	1116	108	50
HR-ICP-MS	CX48-1-689_C2_09	282123	0.01	1.6198	0.0571	0.1588	0.0087	0.641	0.0620	0.0022	534	358	626	9	1017	152	47
HR-ICP-MS	CX48-1-689_C4_01	7301	0.02	1.0266	0.0499	0.1032	0.0077	0.763	0.0778	0.0019	844	314	912	8	1123	99	25
HR-ICP-MS	CX48-1-689_C4_02	93539	0.02	0.9449	0.0345	0.0924	0.0041	0.621	0.0743	0.0016	843	219	942	4	839	89	0
HR-ICP-MS	CX48-1-689_C4_02a	168767	0.01	1.0473	0.0685	0.1063	0.0088	0.719	0.0734	0.0028	733	427	791	9	1014	156	28
HR-ICP-MS	CX48-1-689_C4_02a[1]	5065248	0.01	1.5169	0.0856	0.1443	0.0113	0.784	0.0655	0.0022	631	529	713	11	1132	141	44
HR-ICP-MS	CX48-1-689_C4_03	32071	0.05	0.9578	0.0501	0.0898	0.0061	0.688	0.0787	0.0023	806	315	932	6	817	114	1
HR-ICP-MS	CX48-1-689_C4_04	23286	0.04	1.1159	0.0431	0.1056	0.0056	0.603	0.0777	0.0019	763	272	850	6	1144	98	33
HR-ICP-MS	CX48-1-689_C4_06	14552	0.03	1.0117	0.0514	0.1030	0.0080	0.731	0.0755	0.0024	902	323	947	8	1093	129	18
HR-ICP-MS	CX48-1-689_C4_08	516730	0.00	1.5384	0.0612	0.1400	0.0074	0.833	0.0675	0.0017	524	383	578	8	946	106	45
HR-ICP-MS	CX48-1-689_C4_09	229206	0.01	1.5946	0.0546	0.1487	0.0069	0.619	0.0668	0.0018	624	343	748	7	1168	111	47
HR-ICP-MS	CX48-1-689_C4_10	263783	0.01	1.7593	0.0525	0.1671	0.0087	0.806	0.0680	0.0018	510	330	608	9	1039	113	51
HR-ICP-MS	CX48-1-689_C4_11	1015356	0.00	1.5116	0.0550	0.1336	0.0079	0.789	0.0670	0.0021	626	345	723	8	1103	130	43

Table 5. Isotopic data and apparent ages for uraninite from the Millenniumdeposit (end).

Analytical method	Sample	²⁰⁶ Pb/ ²⁰⁴ Pb	com. Pb*	²⁰⁷ Pb/ ²³⁵ U	±1σ	²⁰⁶ Pb/ ²³⁸ U†	±1σ	R‡	²⁰⁷ Pb/ ²⁰⁶ Pb	±1σ	Apparent ages (±1σ, Ma)**						Disc.††
		(%)									²⁰⁶ Pb/ ²³⁸ U	error	²⁰⁷ Pb/ ²³⁵ U	error	²⁰⁷ Pb/ ²⁰⁶ Pb	error	(%)
HR-ICP-MS	CX48-1-689_C4_12	952994	0.01	1.6981	0.0423	0.1518	0.0044	0.240	0.0673	0.0026	519	267	636	4	1040	160	50
HR-ICP-MS	CX48-1-689_C4_13	351221	0.02	1.5718	0.0573	0.1486	0.0088	0.805	0.0665	0.0021	574	359	692	9	1063	131	46
HR-ICP-MS	CX48-1-689_C4_14	567482	0.02	1.2837	0.0485	0.1181	0.0070	0.641	0.0663	0.0022	757	305	891	7	1207	138	37
HR-ICP-MS	CX48-1-689_C4_15	130609	0.01	1.2715	0.0470	0.1255	0.0066	0.743	0.0683	0.0018	710	296	795	7	1110	110	36
HR-ICP-MS	CX48-1-695_01	146577	0.01	0.5142	0.0169	0.0647	0.0024	0.7135	0.0589	0.0012	404	108	420	2	556	86	27
HR-ICP-MS	CX48-1-695_03	30185	0.02	0.4561	0.0156	0.0647	0.0030	0.1385	0.0598	0.0021	404	100	381	3	567	154	29
HR-ICP-MS	CX48-1-695_04	30496	0.02	1.9615	0.0730	0.1961	0.0104	0.638	0.0811	0.0030	528	454	670	10	1147	143	54
HR-ICP-MS	CX48-1-695_206	131572	0.02	1.3396	0.0190	0.1239	0.0020	0.597	0.0681	0.0008	659	121	740	2	1075	46	39
HR-ICP-MS	CX48-1-695_208	126709	0.02	1.6564	0.0230	0.1384	0.0022	0.829	0.0757	0.0005	835	147	988	2	1081	27	23
HR-ICP-MS	CX48-1-695_210	13744	0.20	0.2977	0.0091	0.0262	0.0010	0.727	0.0756	0.0013	166	59	263	1	1057	70	84
HR-MC-ICP-MS	CX48-1-689_N01	768339	0.01	1.4429	0.0058	0.1317	0.0005	0.956	0.0778	0.0000	631	38	708	1	1074	1	41
HR-MC-ICP-MS	CX48-1-689_N01a	80613	0.01	1.3238	0.0045	0.1220	0.0004	0.952	0.0771	0.0000	660	29	744	0	1075	1	39
HR-MC-ICP-MS	CX48-1-689_N02	52102	0.02	0.9809	0.0090	0.0963	0.0007	0.958	0.0724	0.0001	718	58	835	1	794	8	10
HR-MC-ICP-MS	CX48-1-689_N03	32914	0.02	1.4311	0.0051	0.1300	0.0004	0.945	0.0782	0.0000	657	33	778	0	1108	1	41
HR-MC-ICP-MS	CX48-1-689_N04	410688	0.01	1.2907	0.0030	0.1194	0.0002	0.941	0.0767	0.0000	639	19	716	0	1021	2	37
HR-MC-ICP-MS	CX48-1-689_N05	94532	0.01	1.2772	0.0066	0.1185	0.0005	0.956	0.0765	0.0001	690	42	806	1	1095	3	37
HR-MC-ICP-MS	CX48-1-689_N06	159239	0.01	1.3625	0.0081	0.1253	0.0007	0.959	0.0772	0.0001	668	52	793	1	1103	3	39
HR-MC-ICP-MS	CX48-1-695_101	312661	0.02	1.8589	0.0198	0.1585	0.0017	0.964	0.0835	0.0000	948	127	1066	2	1281	2	26
HR-MC-ICP-MS	CX48-1-695_1C	39046	0.04	0.5363	0.0121	0.0577	0.0012	0.9494	0.0663	0.0001	362	78	435	1	815	8	56
HR-MC-ICP-MS	CX48-1-695_1D	19368	0.07	0.5695	0.0102	0.0637	0.0010	0.7349	0.0642	0.0002	398	65	458	1	747	11	47
HR-MC-ICP-MS	CX48-1-695_1K	42422	0.03	0.9786	0.0105	0.0932	0.0010	0.948	0.0748	0.0001	487	67	558	1	510	8	5
HR-MC-ICP-MS	CX48-1-695_1L	764765	0.03	1.4069	0.0128	0.1247	0.0012	0.954	0.0804	0.0001	593	82	694	1	997	6	41
HR-MC-ICP-MS	CX48-1-695_1M	62855	0.01	1.5130	0.0117	0.1328	0.0011	0.960	0.0810	0.0001	673	75	757	1	1188	4	43
HR-MC-ICP-MS	CX48-1-695_1N	172974	0.02	2.0558	0.0233	0.1728	0.0020	0.965	0.0846	0.0000	1027	149	1133	2	1307	1	21
HR-MC-ICP-MS	CX48-1-695_1P	92440	0.01	1.6124	0.0212	0.1396	0.0018	0.966	0.0821	0.0000	531	135	600	2	995	2	47
HR-MC-ICP-MS	CX48-1-695_1Q	108658	0.02	1.5171	0.0106	0.1331	0.0009	0.960	0.0811	0.0001	507	68	595	1	912	3	44
HR-MC-ICP-MS	CX48-1-695_1R	73853	0.02	1.1533	0.0135	0.1073	0.0012	0.963	0.0765	0.0000	790	86	877	1	1187	2	33
HR-MC-ICP-MS	CX48-1-695_1W	65207	0.02	0.9702	0.0072	0.0899	0.0005	0.922	0.0768	0.0002	1151	46	1098	1	1201	9	4
HR-MC-ICP-MS	CX48-1-695_1Z	43740	0.04	1.0067	0.0094	0.0933	0.0008	0.952	0.0770	0.0001	752	61	860	1	854	4	12
HR-MC-ICP-MS	CX56-3-711_103	834538	0.02	0.9761	0.0403	0.0798	0.0033	0.908	0.0870	0.0001	495	254	689	3	1361	6	64
HR-MC-ICP-MS	CX56-3-711_105	61132	0.01	0.9762	0.0175	0.0773	0.0014	0.797	0.0898	0.0002	480	112	692	1	1421	7	66
HR-MC-ICP-MS	CX56-3-711_109	67566	0.02	1.2778	0.0281	0.1038	0.0021	0.927	0.0876	0.0001	637	179	835	2	1373	6	54
HR-MC-ICP-MS	CX56-3-711_110	78053	0.01	1.3116	0.0243	0.1070	0.0020	0.958	0.0871	0.0001	655	155	850	2	1363	2	52
HR-MC-ICP-MS	CX56-3-711_111	53951	0.01	1.2723	0.0376	0.1049	0.0031	0.944	0.0862	0.0001	642	238	831	3	1343	3	52

* com. Pb (%) = % ²⁰⁶Pb from common lead.

† Measured isotope ratios (except ²⁰⁶Pb/²⁰⁴Pb) and calculated apparent dates are corrected for common lead, as estimated from monitored counts of ²⁰⁴Pb.

‡ R = error correlation coefficient.

§ ²⁰⁶Pb/²³⁸U, ²⁰⁷Pb/²³⁵U, and ²⁰⁷Pb/²⁰⁶Pb ages calculated using equations reported by Ludwig (2000).

†† Disc. = percent discordance.

DL = Detection Limit

Table 6. Pb isotopic compositions and concentrations of Pb and U (ppb) in the leachates from 2 cross-sections of the Manitou Falls Formation sandstones and 1 cross-section of the basement rocks.

Sample	²⁰⁶ Pb/ ²⁰⁴ Pb	²⁰⁷ Pb/ ²⁰⁴ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb age	²⁰⁷ Pb/ ²³⁵ U	²⁰⁶ Pb/ ²³⁸ U	Pb (ppb)	U (ppb)
Manitou Falls Formation sandstones:								
CX43-399	16.64	14.87	0.89	5078	978.32	9.20	0.26	0.01
CX43-522	17.98	14.13	0.79	4896	595.77	6.47	0.11	0.00
CX43-595	31.02	16.10	0.52	4296	8.06	0.11	8.00	14.63
CX44-392	20.20	15.09	0.75	4825	64.13	0.66	0.13	0.05
CX44-516	22.63	15.43	0.68	4693	175.76	1.73	0.21	0.03
CX44-577	25.88	15.01	0.58	4458	20.34	0.26	0.13	0.12
CX48-567.1	29.39	16.60	0.56	4420	84.45	1.07	1.32	0.28
CX48-570.8	30.26	16.63	0.55	4380	31.36	0.40	1.43	0.19
CX49-385	25.90	15.78	0.61	4530	3189.86	41.27	2.38	0.02
CX49-543	15.84	13.72	0.87	5035	111.13	1.11	0.07	0.01
CX49-628	25.95	15.93	0.61	4541	37.93	0.46	1.15	0.56
CX50-400	18.22	15.10	0.83	4972	333.87	2.96	0.24	0.02
CX50-542	22.34	15.55	0.70	4722	95.73	1.03	0.19	0.04
CX50-610	33.89	16.44	0.49	4197	29.17	0.48	0.35	0.21
CX51-381	17.32	14.54	0.84	4990	409.78	3.66	0.15	0.01
CX51-500	17.91	14.36	0.80	4925	77.35	0.70	0.06	0.02
CX51-663	18.75	14.32	0.76	4856	6.98	0.07	0.08	0.25
CX52-392	16.94	15.15	0.89	5080	301.02	2.92	0.10	0.00
CX52-519	17.59	15.14	0.86	5026	584.35	5.00	1.03	0.05
CX52-586	23.62	15.62	0.66	4649	5.09	0.06	0.36	1.59
CX53-1-617.5	30.74	17.55	0.57	4436	219.70	2.64	0.87	0.28
CX53-1-622.5	28.48	16.45	0.58	4452	24.06	0.30	0.57	1.73
CX54-350	16.94	15.00	0.89	5066	968.05	5.53	0.21	0.01
CX54-482	16.15	13.83	0.86	5019	108.11	1.22	0.06	0.01
CX54-548	30.92	16.21	0.52	4311	3.14	0.05	0.07	0.54
CX56-066	24.74	16.33	0.66	4646	523.27	5.51	0.10	0.33
CX56-186.5	26.39	18.12	0.69	4703	32.59	0.34	0.13	0.09
CX56-265	22.72	16.14	0.71	4752	419.18	4.18	0.16	0.08
CX56-355	19.45	16.01	0.82	4962	98.75	0.83	0.12	0.07
CX56-425	19.31	15.87	0.82	4960	149.43	1.30	0.13	0.06
CX56-499.5	21.50	16.46	0.77	4859	88.94	0.82	0.85	0.25
CX56-573	31.46	16.74	0.53	4333	216.17	2.86	0.50	0.20
CX58-371	17.39	14.81	0.85	5011	219.38	1.95	0.11	0.01
CX58-486	18.39	14.91	0.81	4940	198.78	1.76	0.15	0.01
CX58-545	21.18	15.17	0.72	4764	11.72	0.12	0.07	0.14
CX60-391	17.19	14.04	0.82	4951	110.80	1.13	0.08	0.01
CX60-525	18.94	14.37	0.76	4846	25.07	0.24	0.05	0.04
CX60-596	18.64	15.05	0.81	4935	134.36	1.39	0.12	0.05
CX61-358	16.90	14.71	0.87	5042	233.43	2.85	0.12	0.01
CX61-451	18.06	14.91	0.83	4967	244.21	1.92	0.06	0.00
CX61-498	17.24	14.27	0.83	4970	105.23	1.07	0.09	0.02
Basement rocks:								
CX48-593-0	35.94	17.41	0.48	4194	18.90	0.27	6.52	0.16
CX48-625	159.57	22.23	0.14	2219	0.77	0.04	3.35	0.16
CX48-647	144.57	26.85	0.19	2705	2.66	0.11	1.99	5.57
CX48-678	171.05	20.34	0.12	1940	5.16	0.31	2.90	0.19
CX48-688	195.32	23.95	0.12	1995	0.95	0.05	4.22	0.33
CX48-713	516.03	52.17	0.10	1644	0.26	0.02	3.07	0.13
CX48-733	199.00	33.45	0.17	2539	77.03	3.34	3.21	0.17
CX48-852.5	44.48	17.71	0.40	3902	6.28	0.11	0.73	0.12
CX48-856	74.99	19.82	0.26	3272	33.70	0.91	6.42	0.27
CX53-1-629	27.55	16.49	0.60	4505	80.18	0.95	1.62	0.09
CX53-1-657	33.38	17.80	0.53	4336	4.73	0.06	1.52	0.13
CX53-1-664	33.93	16.89	0.50	4235	8.96	0.12	0.40	0.40
CX53-1-705.5	76.46	21.86	0.29	3396	15.36	0.38	4.00	0.19