

Comparison of diagenetic fluids in the Proterozoic Thelon and Athabasca Basins, Canada: implications for protracted fluid histories in stable intracratonic basins

C. Renac, T.K. Kyser, K. Durocher, G. Dreaver and T. O'Connor

Abstract: The Paleoproterozoic Thelon Basin, located on the border between Nunavut and the Northwest Territories of Canada, is a contemporaneous analog of the uranium-rich Paleoproterozoic Athabasca Basin in Canada. Early diagenesis resulted in precipitation of extensive hematite on the surfaces of detrital quartz grains throughout the Thelon Formation and minor hydroxy-phosphate in veins locally. Continued diagenesis then resulted in syntaxial quartz cementation of detrital quartz at 130°C from fluids having ca. 17 wt.% equivalent NaCl, similar to the Athabasca Basin. Cementation of this type is most pronounced in fine-grained sequences in the Thelon Basin. A period of extensive desilicification during continued burial was followed by formation, at ca. 200°C, of peak-diagenetic illite having Ar–Ar ages of ca. 1400–1690 Ma in the Thelon Formation. This illite was associated with fluids with $\delta^{18}\text{O}$ and δD values of ca. 6‰ and –50‰, respectively, similar to those during peak diagenesis of the Athabasca Basin. Although the timing, salinity, and isotopic composition of the peak-diagenetic fluids in the Thelon and Athabasca Basins are similar, the peak-diagenetic mineral assemblage in the Athabasca Formation is dickite and illite, with minor dravite and goyasite rather than simply illite. Consequently, the fluids at peak diagenesis, which in the Athabasca Basin are synchronous with formation of world-class unconformity-type uranium deposits, had different compositions in each basin. Post-peak diagenesis in the Thelon Basin was quite distinct from that in the Athabasca Basin in that illite was replaced in the central portion of the basin by K-feldspar and then sudoite, which crystallized from saline brines at ca. 1000 Ma and 100°C. Evidence for later infiltration of these brines is absent in the Athabasca Basin, although uranium mobilization at ca. 900 Ma from fluids having the same characteristics as those at peak diagenesis was pronounced in the Athabasca Basin. Recent incursion of meteoric waters along reactivated structures into the Athabasca Basin has variably affected hydrous and uraniferous minerals, but evidence for this is lacking in the Thelon Basin. The Thelon Basin reflects less intensive fluid–rock interaction in its early history than that recorded in the basal units of the Athabasca Basin.

Résumé : Le bassin de Thelon (Paléoprotérozoïque), situé sur la frontière entre le Nunavut et les Territoires du Nord-Ouest du Canada, est l'analogue contemporain du bassin riche en uranium d'Athabasca (Paléoprotérozoïque), au Canada. Une diagenèse précoce a fait précipiter une grande quantité d'hématite sur les surfaces des grains de quartz détritiques à travers toute la Formation de Thelon et, localement, une faible quantité d'hydroxy-phosphate dans les veines. La poursuite de la diagenèse a ensuite résulté en la cimentation quartzique par surcroissance du quartz détritiques à 130 °C à partir de fluides qui avaient environ 17 % (poids) d'équivalent NaCl, semblable au bassin d'Athabasca. La cimentation de ce type est le plus prononcée dans les séquences à grains fins du bassin de Thelon. Une période de désilicification extensive, au cours d'un enfouissement continu, a été suivie de la formation d'illite au sommet de la diagenèse, à environ 200 °C; les âges Ar–Ar de l'illite sont d'environ 1400–1690 Ma dans la Formation de Thelon. Cette illite était associée à des fluides ayant des valeurs de $\delta^{18}\text{O}$ et δD d'environ 6‰ et –50‰ respectivement, semblables à celles présentes durant le sommet de la diagenèse du bassin d'Athabasca. Bien que la séquence, la salinité et la composition isotopique des fluides au sommet de la diagenèse dans les bassins de Thelon et d'Athabasca soient semblables, l'assemblage minéralogique du sommet de la diagenèse dans la Formation de l'Athabasca consiste en dickite et en illite avec de la dravite et de la goyasite plutôt que simplement de l'illite. Par conséquent, les fluides au sommet de la diagenèse, lesquels dans le bassin d'Athabasca sont synchrones à la formation de dépôts d'uranium de classe mondiale et de type discordance, ont des

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compositions différentes dans chaque bassin. Dans le bassin de Thelon, la diagenèse après le sommet a été assez différente de celle dans le bassin d'Athabasca en ce que l'illite dans la partie centrale du bassin a été remplacée par un feldspath-K puis par de la sudoïte, qui a cristallisé à partir de saumures vers 1000 Ma et à 100 °C. Il n'existe aucune preuve d'une infiltration postérieure de ces saumures dans le bassin de l'Athabasca, bien que la mobilisation de l'uranium, vers 900 Ma et à partir de fluides ayant les mêmes caractéristiques, ait été forte dans le bassin de l'Athabasca. Une récente incursion d'eaux météoriques dans le bassin de l'Athabasca, le long de structures réactivées, a affecté les minéraux hydratés et uranifères à divers degrés, mais il n'y a pas d'évidence de cela dans le bassin de Thelon. Le bassin de Thelon reflète une interaction fluide-roche moins intense au début de son histoire que celle enregistrée dans les unités à la base du bassin de l'Athabasca.

[Traduit par la Rédaction]

Introduction

The Paleoproterozoic Thelon Basin straddles the border between Nunavut and the Northwest Territories of Canada (Fig. 1). It is a contemporaneous analog of the uranium-rich Athabasca Basin in Canada and Kombolgie Basin in Australia. Sedimentary deposition in all three basins started at ca. 1750–1720 Ma (Page 1988; Miller et al. 1989; Kotzer and Kyser 1995; Kyser et al. 2000), with rapid regional uplift after the Proterozoic Trans-Hudson Orogen in Canada and Baramundi Orogen in Australia at ca. 1800–1900 Ma (Kyser et al. 2000). The Athabasca and Kombolgie basins host world-class unconformity-type uranium deposits that formed as a result of long-lived diagenetic and hydrothermal fluid systems (Gustafson and Curtis 1983; Wilde et al. 1989; Kotzer and Kyser 1995). In contrast, no large uranium deposits have been discovered within the Thelon Basin, despite similarities in the sedimentological and evolutionary record of these three basins (e.g., Miller 1995). The timing and character of fluid events in the Thelon Basin are poorly constrained and the similarities between it and the aforementioned Proterozoic basins are drawn primarily on general sedimentological rather than diagenetic features. In this study, we present the fluid evolution of the Thelon Basin from detailed petrographic, geochemical, and geochronological results and use these in conjunction with data from the Athabasca Basin (Kotzer and Kyser 1995) to assess similarities and differences in the hydrologic and diagenetic evolution of ancient sedimentary basins.

Geologic considerations

Stratigraphy and sedimentology

The Athabasca Basin is located within northern Saskatchewan and the Thelon Basin is located along the Nunavut–Northwest Territories boundary. Both are located in the Slave and Churchill geological provinces (Fig. 1). These basins are two of a series of spatially and temporally related Paleo- to Mesoproterozoic intracratonic basins floored by Archean and Paleoproterozoic basement rocks. Other such basins include those in the Pine Creek geosyncline in the Northern Territory, Australia, and remnants of basins in Nunavut (i.e., Baker, Borden, and Hornby Bay basins). The Canadian basins formed following the Trans-Hudson Orogen and consist of thick sequences of mature quartz sandstone, conglomerate, siltstone, and shale (e.g., Athabasca Group and Thelon formations). These units lie unconformably on paleoregoliths on Archean and Paleoproterozoic crystalline basement rocks.

The geology and sedimentology of the Athabasca Basin have been the subject of several studies because of the uranium

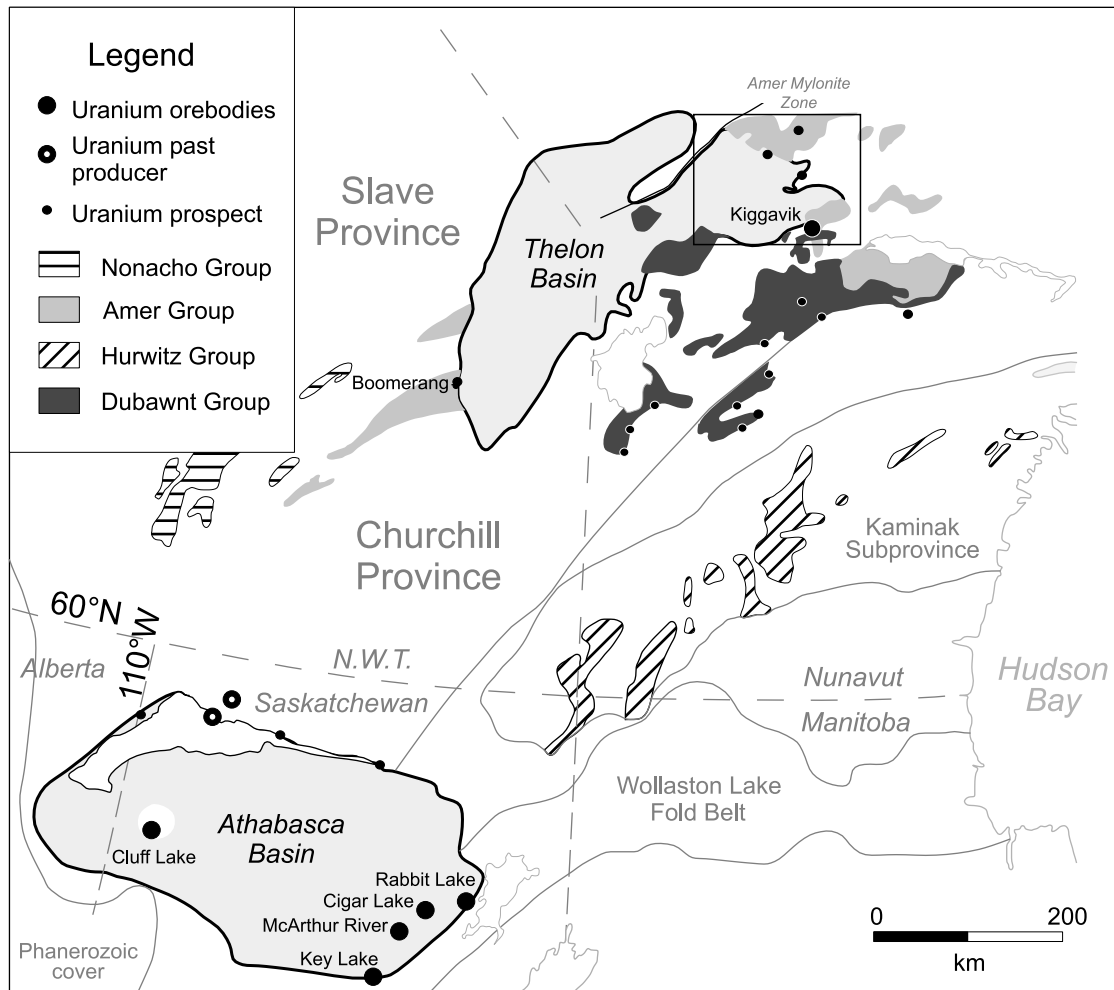
deposits it hosts. Summaries of these studies and the deposits can be found in Ramaekers (1990), Kotzer and Kyser (1995), Ruzicka (1996) and Kyser et al. (2000).

A large number of reconnaissance projects around the area of the Thelon Basin over the past 40 years (e.g., Wright 1955; Donaldson 1965; LeCheminant et al. 1983; Miller 1995) have resulted in an inconsistent array of basin and basement rock unit terminology and correlations. The eastern portion of the Thelon Basin, which will be referred to as the Thelon sub-basin (Fig. 2), is the focus of this study and has been an active area for mineral exploration. Descriptions of the Thelon Formation in the eastern and western sub-basins are similar, implying they have had similar diagenetic histories. The vast western portion of the basin is a game sanctuary and is not easily accessible.

Basement rocks in the Thelon sub-basin include a series of Archean metapelites and gneisses of the Woodburn Lake Group, Aphebian metaigneous, metasedimentary, and sedimentary rocks of the Amer and Baker Lake groups, and Late Aphebian sedimentary and volcanic rocks of the Wharton Group (Fig. 2). The relationship between the Amer and Baker Lake groups has yet to be resolved, although the former may be either older or related to the lowest units in the latter. The overlying Helikian Thelon Formation belongs to the Barrenland Group and comprises basal conglomerates of variable thickness and fluvial to near-shore quartz arenite units up to 2 km thick (Donaldson 1969; Gall et al. 1992). The older Paleoproterozoic units in this area are part of the Dubawnt Group (Fig. 2), whereas all of the Paleoproterozoic units compose the Dubawnt Supergroup of Gall et al. (1992). A paleoregolith, consisting of a paleosol horizon variably preserved on the sub-Thelon Formation unconformity, marks the boundary between the Thelon Formation and basement rocks (Gall 1994). The regolith and basal units of the Thelon Formation contain cements of possible diagenetic phosphate minerals with U–Pb ages of 1720–1760 Ma (Miller et al. 1989). These ages are considered minimum ages for Thelon Formation sedimentation (Miller 1995). The maximum age for formation of the basin is constrained by the age of emplacement of fluorite-bearing granites into the Amer Group at ca. 1753 Ma (Miller 1995). Later events preserved in the eastern portion of the basin include Mackenzie diabase dikes at ca. 1270 Ma (LeCheminant and Heaman 1989), deposition of Ordovician limestones, and Quaternary glacial deposits.

The Thelon Formation in the western portion of the basin is capped by thin basaltic flows of the Kuungmi (or Sanctuary) Formation. These volcanic rocks are stratigraphically overlain

Fig. 1. Location and regional geology of the area around the Athabasca and Thelon basins, Canada. Shown are Churchill and Slave provinces, major structures and units, locations of major uranium ore bodies in the Athabasca Basin and Kiggavik and Boomerang prospects in the Thelon Basin. Area in rectangle is the Thelon sub-basin shown in Fig. 2.



by carbonates and evaporites of the Helikian Lookout Point Formation (Fig. 2). The Kuungmi Formation flows attest to late igneous activity during basin evolution, and the Lookout Point Formation to evaporitic brines that may have circulated and diagenetically altered the Thelon Formation sediments. Hiatt et al. (1999) showed that sands deposited in upper shoreface environments occur at multiple stratigraphic horizons within the Thelon Formation of the eastern sub-basin. This further increases the likelihood of seawater-derived pore fluids in the Thelon Basin. These late marine units in the Thelon Basin are also strikingly similar to the stratigraphy recorded in the uranium-rich Athabasca Basin (Ramaekers 1990).

The Thelon Formation in the southeastern part of the sub-basin was divided into three units by LeCheminant et al. (1983). The lowermost unit is dominated by conglomerates and associated sandstones, siltstones, and mudstones. This unit has been interpreted as reflecting a continental setting of "active tectonism." The conglomerates contain abundant clasts of the Pitz Formation volcanic rocks (Fig. 2), and all lithologies in this unit contain detrital feldspar and are completely cemented by quartz. Paleocurrents are predominantly northward-directed. This unit is referred to as the pre-Thelon Formation Sandstone unit throughout this paper. This lowermost

unit is overlain by conglomerates rich in basement quartzite and sandstone clasts. In this overlying unit, the transition to sandstone takes place up section, where trough cross-bedding and fining-up cycles are observed. Rare angular unconformities between the pre-Thelon Formation Sandstone and the basal Thelon Formation suggest either synsedimentary faulting or that the pre-Thelon Formation sandstone unit predates the Thelon Formation. The overlying unit is not extensively quartz cemented, but rather "clay cemented" lacking preserved detrital feldspars and only rare clasts of the Pitz volcanics (LeCheminant et al. 1983). Paleocurrents are generally westward directed. This unit is more typical of the Thelon Formation throughout the basin. An upper unit consists of sandstone interpreted as formed from aeolian processes. Paleodirections suggest that dominant winds were from the east.

Economic considerations

The Athabasca Basin hosts several unconformity-type uranium deposits. Fluid histories of these have been documented using several geochemical techniques (e.g., Pagel and Jaffrezic 1977; Pagel et al. 1993; Philippe et al. 1993; Wilson and Kyser 1987; Kotzer et al. 1992; Kotzer and Kyser 1995). These studies show four major tectono-fluid events that have

influenced the genesis of these deposits: (1) an early diagenetic event at 1700–1600 Ma, which precipitated hematite and quartz overgrowths at ca. 150°C; (2) high-temperature (>200°C) burial diagenesis with oxidizing and saline fluids at 1600–1450 Ma; (3) fluid mobilizing uranium and recrystallizing Fe-oxides at ca. 900 Ma; and (4) uplift and faulting related to Phanerozoic movement.

The Thelon Basin is spatially and temporally related to the Athabasca Basin and is host to two areas of uranium mineralization, the Kiggavik deposit and Boomerang Lake prospect (Fig. 1). The Boomerang Lake prospect has a complex metal inventory (Gandhi 1989), not unlike those of typical complex unconformity-type uranium deposits in the Athabasca Basin. This prospect is limited in extent, is hosted by the Thelon sandstones and in graphitic basement rocks, and has an apparent age of ca. 1300 Ma (Gandhi 1989). In contrast, the Kiggavik deposit is hosted by late Archean and Paleoproterozoic basement rocks that lack any graphite (Fuchs and Hilger 1989). The Thelon Formation has been removed from the area around Kiggavik by erosion, but it is exposed about 2 km north of the deposit. The alteration assemblage and the few ages of ca. 1300 Ma are similar to those observed in the uranium deposits of the Athabasca Basin (e.g., Kotzer and Kyser 1995). If uranium mineralization in these two areas is unconformity type, as some of the preliminary studies would suggest, then the Thelon Basin could potentially host several economic concentrations of uranium that formed during its fluid evolution.

Sampling and methodology

Samples were taken from seven drill holes of variable depths from the Thelon sub-basin (Figs. 2, 3). Petrography, cold cathodoluminescence, scanning electron microscopy (SEM), and electron-probe microanalyses were used to decipher the mineral paragenesis of the sediments. Three size fractions were separated (>5 µm, 5–2 µm, and <2 µm) from selected samples by settling after gentle crushing with deionized water and ultrasonic dispersion. X-Ray diffraction (XRD) patterns were obtained using a Siemens diffractometer utilizing Cu K α radiation (35 kV, 15 mA) and two fixed slits (entrance 1°, exit 0.1°). Disoriented and oriented patterns were obtained using a quartz internal standard (step: 0.025°, time: 8 s). Clay proportions were calculated using Newmod simulations (Reynolds 1985) and microprobe results. Illite crystallinity was estimated after background subtraction using peak width at half of maximum intensity of the 00L peaks.

Microthermometric measurements on doubly polished thick sections were obtained using a U.S. Geological Survey heating-freezing stage. For aqueous inclusions, the initial and final melting temperatures of ice and total homogenization temperatures (Th) have errors of ± 0.2 and 2°C respectively. Homogenization temperatures were assumed to approximate trapping temperatures, because pressure corrections would be less than ca. 40°C for units buried to depths of less than a few kilometres. The microthermometric data provide information on fluid density, salinity, and trapping temperatures (Roedder 1984; Zhang and Frantz 1987).

Ar–Ar ages were obtained from irradiated whole-rock chips (1 cm² by 0.5 mm thick) and mineral separates (<20 µm fractions) at the Queen's University Ar-laser laboratory. The

Ar in illite and K-feldspar in pore spaces in the whole-rock chips were released by spot fusion using an Ar laser and focussed spots. Pore spaces larger than 30 µm were sampled several times using this technique to ensure consistent ages of the minerals. Mineral separates consisted of pure illite or mixtures of feldspar and chlorite. The small size of the feldspar and its association with chlorite precluded separation. Step heating of the separates was done in 10 steps from 0.25 to 4 W. All release patterns were disturbed because of recoil effects. As a consequence, total fusion ages for the separates are used, although they are comparable to the plateau ages for the higher temperature steps.

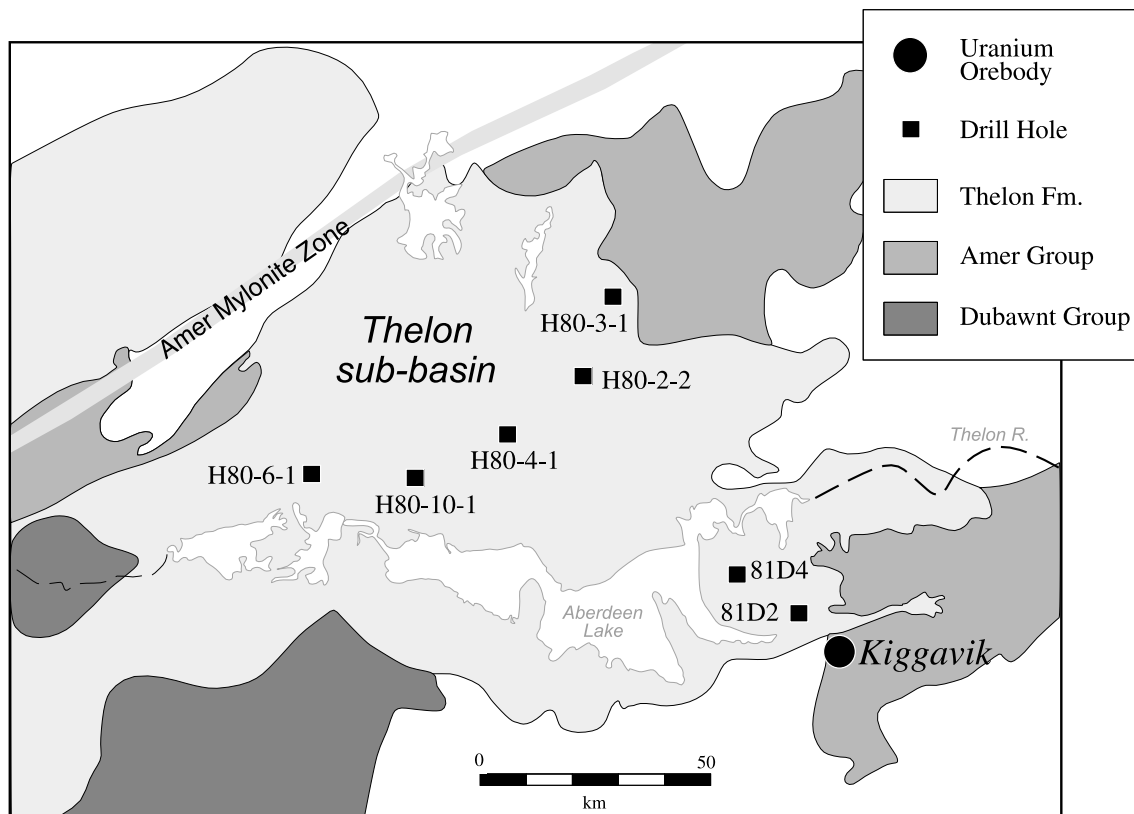
Stable-isotope compositions were measured at Queen's University using the conventional BrF₅ technique outlined by Clayton and Mayeda (1963) for oxygen and by Kyser and O'Neil (1984) for hydrogen. When mixtures of clays were analyzed, isotopic compositions of end members were determined by correlation lines extrapolated using XRD relative proportions and water contents. Carbon and O isotopic measurements were done using the method of McCrea (1950).

Isotopic compositions of fluids were calculated using the following oxygen isotope fractionation equations: dolomite–water of Sheppard and Schwarcz (1970); quartz–water of Kawabe (1978), Land and Dutton (1978), and Taylor (1974); kaolinite and illite – water of Land and Dutton (1978); hematite–water of Becker and Clayton (1976); chlorite and K-feldspar – water of Savin and Lee (1988). The following hydrogen isotope fractionation equations were used: kaolinite–water of Liu and Epstein (1984); chlorite–water of Marumo et al. (1980); illite–water reactions were approximated using the smectite–water fractionation equation of Yeh (1980).

Mineral paragenesis in the Thelon Basin

Samples of drill core from the central portion of the Thelon sub-basin are from flat-lying sandstones of the Thelon Formation. These consist of medium- to fine-grained detrital quartz surrounded by a rim of syntaxial, syncompaction quartz overgrowths and up to 20% clay interstitial filling (Fig. 3). Throughout the Thelon Formation, the sandstones have rare clasts of quartzite, schist, gneiss, and fine-grained phosphate. Basal sandstone units, which are from the edges of the sub-basin, tend to be fine-grained and quartz-cemented, whereas those stratigraphically higher tend to alternate between fine-grained, silicified sandstones and coarser, friable sandstones. Samples from near the southern and northern edges the sub-basin are quartz-cemented basal conglomerates having clasts of the Pitz Formation, quartzite, and large microcline grains. At the base of northeast edge, most drill holes intersect metamorphosed Amer Group rocks capped by a paleoregolith, over which sits a flat-lying arkosic sandstone unit, which is part of the pre-Thelon Formation Sandstone unit. This unit consists of detrital feldspar and quartz, with isopachous quartz cements indicative of vadose zone quartz precipitation surrounding detrital grains and later quartz overgrowths and even later pore-filling authigenic clay minerals. The sandstone is extremely quartz-cemented and fine-grained and most likely records significant diagenesis prior to deposition of the Thelon Formation.

Fig. 2. General geology and location of drill holes for the Thelon sub-basin. Geology includes surrounding metamorphic (Amer group), plutonic (Dubawnt Supergroup), and volcanic (Pitz Formation) units. Stratigraphic relations of basement units of the Dubawnt Supergroup and sedimentary lithologies are also shown along with ages. Modified from Gall et al. (1992). Fm., Formation; Gp., Group.



Magmatic and sedimentological sequence of the Thelon sub-basin

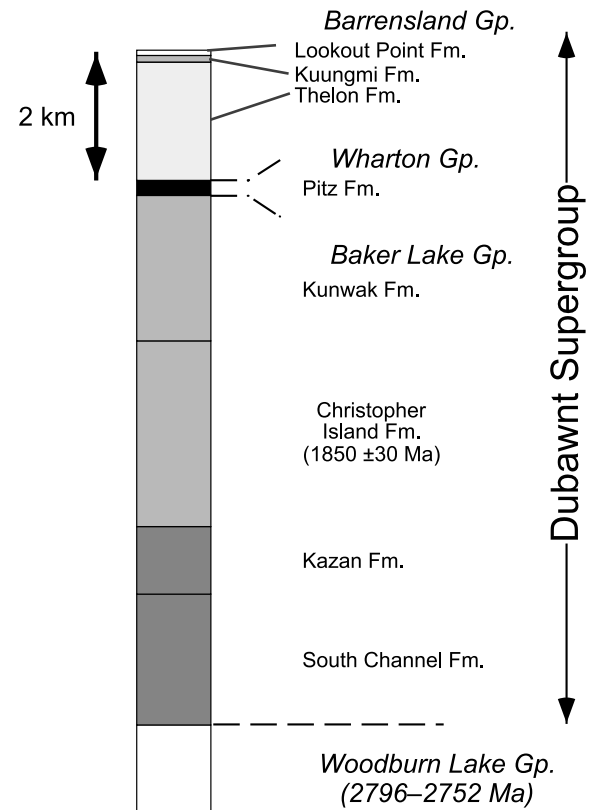
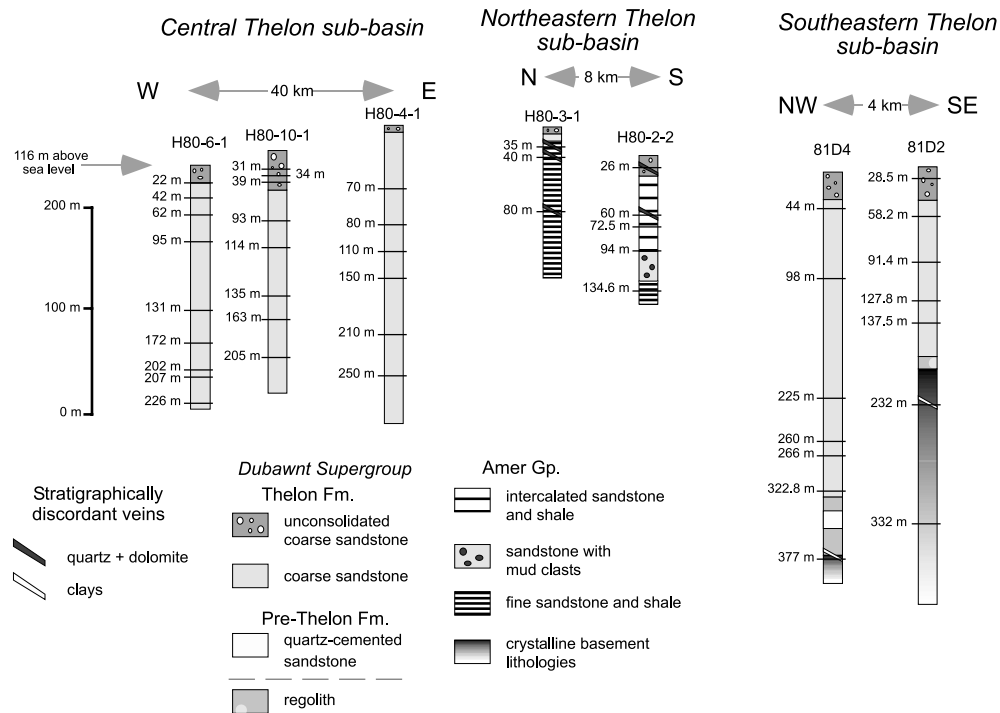


Fig. 3. Geology and stratigraphy of drill cores (H80-6-1, H80-10-1, H80-4-1, H80-3-1, H80-2-2 and 81-D-4, 81-D-2) from the central, northeastern and southeastern Thelon sub-basin, with current altitudes (metres above sea level) and depths sampled. Locations of the drill cores are shown in Fig. 2.



The regolith that directly overlies the Archean and Paleoproterozoic basement lithologies ranges in thickness from 5 to 30 m (Fig. 3). It appears to be primarily accreted, consisting of rounded quartz pebbles in a matrix of quartz, feldspar, lithic fragments, and lesser illite and kaolinite. Rare, millimetre-sized veins within the regolith are filled with quartz, kaolinite, and illite or quartz and dolomite.

Petrographic examination of thin sections reveals a complex history of post-sedimentation events recorded in the Thelon Formation (Figs. 4, 5). Basinwide alteration events can be divided into an early diagenetic phase, peak diagenesis, and late diagenetic event. Less extensive events are represented by small veins and will be referred to as local events.

In the Thelon Formation, the only detrital phases preserved are rare clasts of metamorphic rocks, occasional zircon and muscovite grains, fluorapatite clasts (P0), and abundant detrital quartz (Q0). The detrital quartz, which normally comprises > 95% of the Thelon Formation, occasionally contains inclusions of apatite, rutile, and zircon. In the pre-Thelon Formation Sandstone, detrital microcline (F0) is still preserved as a result of early quartz cementation (Fig. 4). Most detrital quartz grains are between 250 and 500 µm in diameter and well rounded, indicating significant transport and sorting of provenance material.

Early diagenesis

Early diagenesis is manifested by fine veneers of reddish-brown or black iron oxides (H1) and syntaxial, syncompaction quartz overgrowths (Q1) on detrital quartz grains. Petrographic observations show that these oxide veneers are composed of platy hematite of 1 µm to 10 µm in diameter (H1; Fig. 5A). Ultraviolet optical microscopy reveals the presence of detrital

feldspar ghosts in the present pore space in the sandstones, although these spaces are now filled with quartz and illite. The feldspar ghosts are surrounded by ghosts that resemble relict quartz overgrowths, indicating that at least some of the early quartz cementation was prior to feldspar dissolution. Local variations include areas that lack quartz overgrowths but display specular hematite (H2) and associated neoformed hydroxy-phosphate cement (P1; Fig. 5B). Small, millimetre-wide quartz veins containing euhedral quartz (Q2) along the margins and later minerals in the center occur locally in the pre-Thelon Formation Sandstone and basal Thelon formations. These crosscut the quartz overgrowths, but are cut by small veinlets of minerals produced during peak diagenesis.

Peak diagenesis

A period of silica dissolution occurred prior to precipitation of illite in pore spaces during peak diagenesis (Figs. 4, 5B). Illite crystals are typically < 10 µm (I1 and I2) and are distributed radially around quartz grains or remnant quartz overgrowths (Figs. 5C, 5D). There are two generations of illite that occur as pore fillings, but they cannot be distinguished petrographically and, as will be discussed, only on the basis of Ar–Ar ages, stratigraphic position, and stable isotopic compositions. In the lowermost sandstone units, illite has a higher crystallinity (width at half-peak height from 0.3° to 0.7° 2θ on 001) than in upper units (0.6°–0.9° 2θ on 001) with local interlayered clays also present in the upper units (illite/smectite is R₃ ordered with < 10% smectite). Throughout the Thelon Formation, XRD patterns on randomly oriented preparations indicate that the basal units of the pre-Thelon Formation Sandstone and Thelon Formation

Fig. 4. Paragenetic sequence of minerals for the Thelon Formation for early, peak, and late diagenesis. Temperatures of various events are based on fluid inclusion or crystallinity measurements and ages on U–Pb, Ar–Ar, or interpolation (see text for discussion).

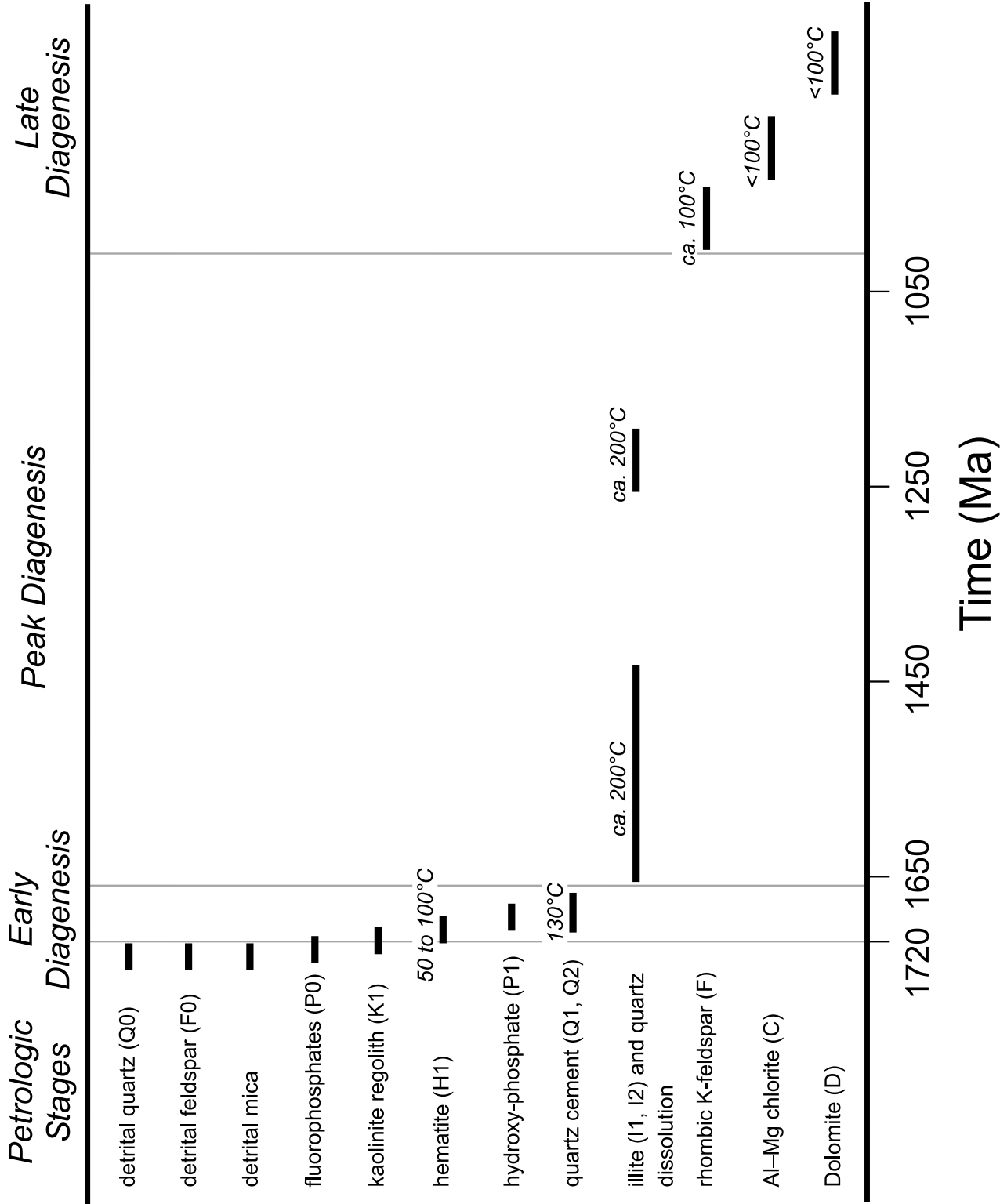
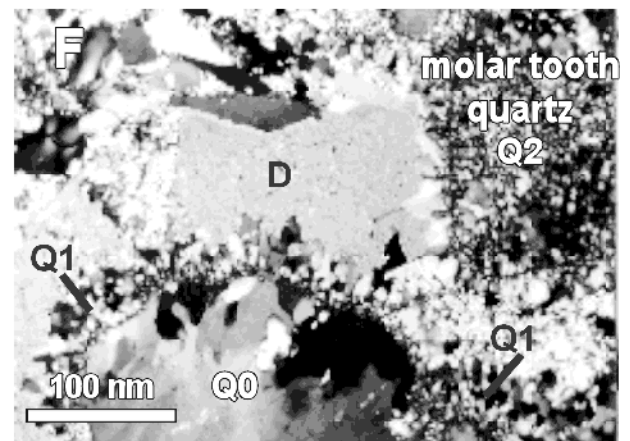
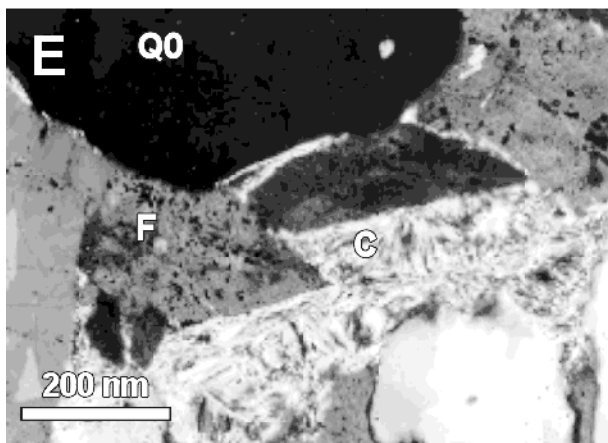
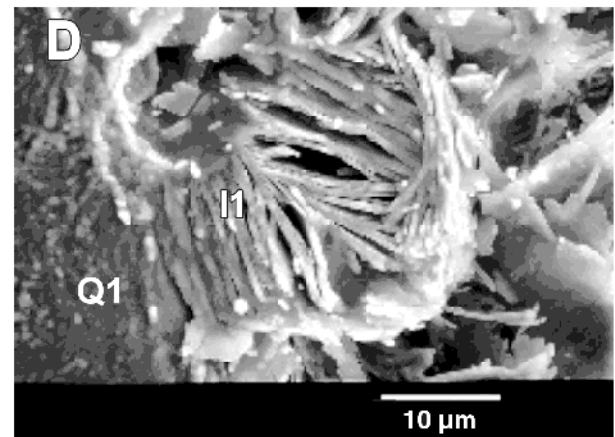
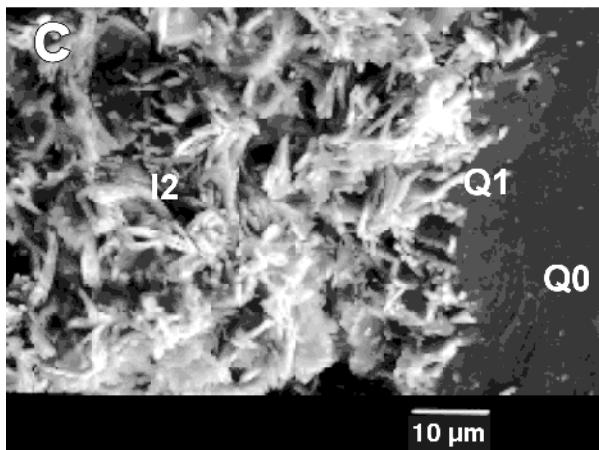
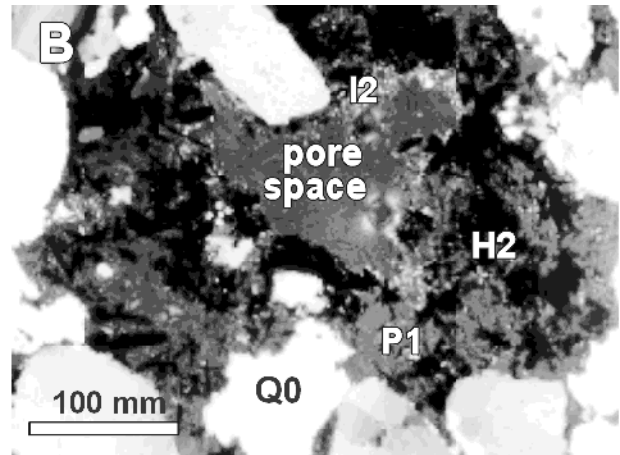
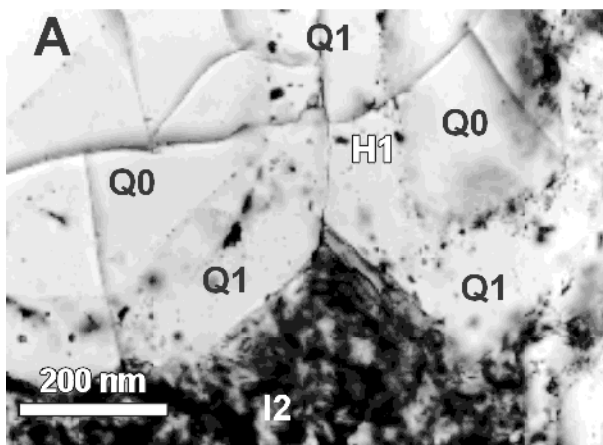


Fig. 5. Microphotographs of thin sections and chips observed under the scanning electron microscope (SEM) of samples representing characteristic aspects of the paragenetic sequence in the Thelon Formation as presented in Fig. 4. (A) Microphotograph (plain light) of a quartz-cemented zone in sample H80-4-1-250m, showing clear detrital quartz (Q0), microscopic hematite (H1) followed by quartz overgrowths (Q1) containing numerous liquid-vapour fluid inclusions and illite (I2). (B) Microphotograph (crossed polars) of phosphatic sandstone near a vein filled with phosphate (P1) and acicular hematite (H2; sample H80-4-1-110 m). Rounded quartz grains (Q0) are cemented with phosphate and hematite and residual spaces are partially filled with illite (I2). (C) SEM photo of quartz grain (Q0) with dissolved quartz overgrowths (Q1) and authigenic illite (I2) from sample H80-2-2-26 m. (D) SEM photo of sample H80-4-1-250m showing radial aspect of illite flakes (I1) growing on quartz overgrowth surface (Q1). (E) Microphotograph (crossed polars) of quartz grains (Q0), rhombic K feldspar (F) and radial Al-Mg chlorite (C) sample H80-6-1-62m. (F) Thin section (polarized light) of sample 81-D-2-137.5m showing detrital quartz grains (Q0) surrounded by euhedral quartz cement (Q1) and molar tooth quartz in veins (Q2). Pore spaces are filled with late dolomite (D).



have only $2M_1$ illite, whereas some of the stratigraphically higher units of the Thelon Formation have mixtures of $2M_1$ and $1M$ polytypes of illite. High-ordered illite polytypes ($2M_1$) suggest precipitation at a higher temperature than the $1M$ polytype. The preservation of detrital muscovite in the basal sandstones and the presence of the high-ordered $2M_1$ illite polymorph are consistent with fluid temperatures near 200°C (Inoue and Utada 1983). Conversely, the low-ordered $1M$ illite polytype found in a few samples from the higher sequences may correspond to partial retrograde alteration of earlier formed $2M_1$ illite. Electron microprobe analyses of 10 samples each of high and low crystallinity illites indicate less K in the latter, which is further evidence of retrograde alteration of the more crystalline clay ($I1_{\text{average}}$: $((\text{Si}^{4+}_{7.05} \text{Al}^{3+}_{0.95})(\text{Al}^{3+}_{3.65} \text{Fe}^{3+}_{0.2})(\text{K}^{+}_{1.4} \text{Ca}^{2+}_{0.44})\text{O}_{22}\text{OH}_4)$; $I2_{\text{average}}$: $((\text{Si}^{4+}_{6.62} \text{Al}^{3+}_{1.38})(\text{Al}^{3+}_{3.45} \text{Fe}^{3+}_{0.44})(\text{K}^{+}_{1.28} \text{Ca}^{2+}_{0.42})\text{O}_{22}\text{OH}_4)$).

Late diagenesis

Late diagenesis is characterized by widespread rhombic K-feldspar (F: $((\text{Si}^{4+}_{2.999} \text{Ti}^{4+}_{0.001})(\text{Al}^{3+}_{1.01})(\text{Fe}^{3+}_{0.03} \text{K}^{+}_{0.96} \text{Ca}^{2+}_{0.003})\text{O}_8)$; $n = 8$) and paragenetically later radial chlorite (C) in the uppermost sandstones of the central Thelon sub-basin (Figs. 4, 5E). The K-feldspar is poorly ordered, consistent with temperatures of precipitation of $< 150^\circ\text{C}$ (Kastner and Siever 1979). Chemical compositions and XRD patterns indicate that the chlorite is di-trioctahedral (polytype $1M_{\text{IIB}}$), having an Mg-rich composition (C_{average} : sudoite: $((\text{Si}^{4+}_{6.98} \text{Al}^{3+}_{1.02})(\text{Al}^{3+}_{5.83} \text{Fe}^{3+}_{0.17} \text{Mn}^{2+}_{0.05} \text{Mg}^{2+}_{3.37})(\text{Na}^{+}_{0.01} \text{K}^{+}_{0.3} \text{Ca}^{2+}_{0.05})\text{O}_{28}\text{OH}_8)$). In samples with diagenetic feldspar and chlorite, or just chlorite, the quartz overgrowths are completely dissolved and the detrital quartz grains are unsupported. The feldspar and chlorite occur as pore fillings after the overgrowths have been dissolved, and feldspar normally rims the pores and is variably replaced by the later chlorite. Within sandstone units containing feldspar and chlorite, these minerals are dominant in zones that are desilicified, whereas illite (I1) is preserved in the more silicified beds. Miller and LeCheminant (1981) suggested on the basis of outcrop samples that chlorite and feldspar reflected distinct stratigraphic units within the Thelon Formation because there was an abrupt change in mineralogy of the sandstones. However, there is no abrupt change in mineralogy of the sandstones with depth and, as will be discussed, the ages of the feldspar are substantially younger than those of the illites.

Millimetric quartz veins (Q2) crosscut the basal unit of the Thelon and the pre-Thelon Formation Sandstone. Some of the veins contain only euhedral quartz and are cut by small veins of illite (I1). Some veins contain euhedral quartz along the edges, ferric oxides, and coarse-grained dolomite (D) in the centers (Figs. 4, 5F). The dolomite precipitated during extension of existing quartz veins and are probably representative of fluid events associated with basement reactivation. Fluids responsible for dolomite precipitation partially dissolved the euhedral quartz. Cathodoluminescence shows that the dolomite crystals within the veins have Fe-rich cores, ca. 1.5 wt.% FeO by electron microprobe.

Timing of fluid events

The timing of specific fluid events can be deduced from

the radiometric ages of diagenetic minerals of known paragenesis. The age for the onset of sedimentation in the basin is ca. 1720 Ma (Miller 1995), based on the youngest U–Pb ages of phosphates, which presumably correspond to the hydroxy-apatite (P1) in our paragenesis (Fig. 4). The oldest ages for the phosphate clasts are 1750 Ma, which correspond to pre-Thelon phosphate sedimentation in lakes (P0; Kyser et al. 2000).

For later K-bearing phases such as illite and K-feldspar, Ar–Ar ages can be used to delineate fluid events (e.g., Harrison 1990). Release patterns for illite separates and mixtures of feldspar and chlorite are disturbed (Fig. 6A) because of loss of Ar from recoil. However, integrated ages and the ages from the highest temperature steps are comparable (Fig. 6A; Table 1). Ar–Ar integrated ages of fusion spots in illite produced during peak diagenesis from the basal units and the pre-Thelon Formation Sandstone range from 1420 to 1690 Ma (Table 1; Fig. 6B). In contrast, those from stratigraphically higher areas from the central portion tend to have younger ages from 1175 to 1588 Ma (Fig. 6A, 6B). The distinction in Ar–Ar ages and stratigraphic position of these illites suggest that they closed to Ar diffusion at different times. Consequently, the older illites are referred to as I1 and the younger as I2 to distinguish them in the paragenesis (Fig. 4). There is no apparent difference in their petrographic relations as both are pore-filling.

The Ar–Ar fusion ages of mica from a quartzite clast from the basal conglomerate range from 1933 to 2149 Ma (Fig. 6B). The mineralogy and ages of micas contained in this clast are similar to the age of the Baker Lake Group (Miller 1995), the probable source of the clast. Early quartz cementation is extensive and results in isopachous cements, preserved ages in clasts, and relatively old Ar–Ar ages. These cements are manifest in the pre-Thelon Formation sandstone units and basal conglomerates of the Thelon Formation and indicate that these units acted as major aquitards relatively early in the evolution of the basin, because the cements filled all available pore space. In contrast, stratigraphically higher units rarely contain preserved detrital micas, have illite that grew into existing pore spaces during peak diagenesis, and have younger illite Ar–Ar ages, consistent with these acting as variable aquifers during peak diagenesis in the basin.

Authigenic K-feldspars have integrated Ar–Ar ages between 880 and 1155 Ma (Figs. 6A, 6B), with most having ages substantially younger than the illite. Those feldspars having the oldest ages are either mixed with illite or have the highest amounts of atmospheric contamination (Table 1) and most likely reflect aberrantly old ages resulting from loss of ^{39}Ar due to recoil (e.g., Harrison and McDougall 1982). The relatively young ages and distinct paragenesis of the feldspars are evidence that they record a distinct temporal and compositional change in the fluids that affected the basin during late diagenesis. For comparison, there is no significant amount of authigenic feldspar or chlorite in the Athabasca Basin.

Fluid inclusion petrography and microthermometry

Primary fluid inclusions in quartz overgrowths (Q1) from samples in drill core from the central and western Thelon sub-basin have average homogenization temperatures ca.

Fig. 6. (A) Ar–Ar ages for fraction of ^{39}Ar released during step-heating of illite and feldspar–illite–chlorite separates from drill cores from the central portion of the Thelon sub-basin. (B) Histograms of integrated Ar–Ar ages obtained by fusion of individual spots in pore spaces of known mineralogy for samples from the Thelon sub-basin. I1 and I2 illite from central (H80-10; H80-6; H80-4), northeast (H80-2-2) and southwestern (81-D-4 and 81-D-2) portions from the Thelon sub-basin are in the left diagram, and feldspar–chlorite and feldspar–illite mixtures (H80-10 and H80-6) on the right. Drill core from which the samples were collected are also indicated (data from Table 1).

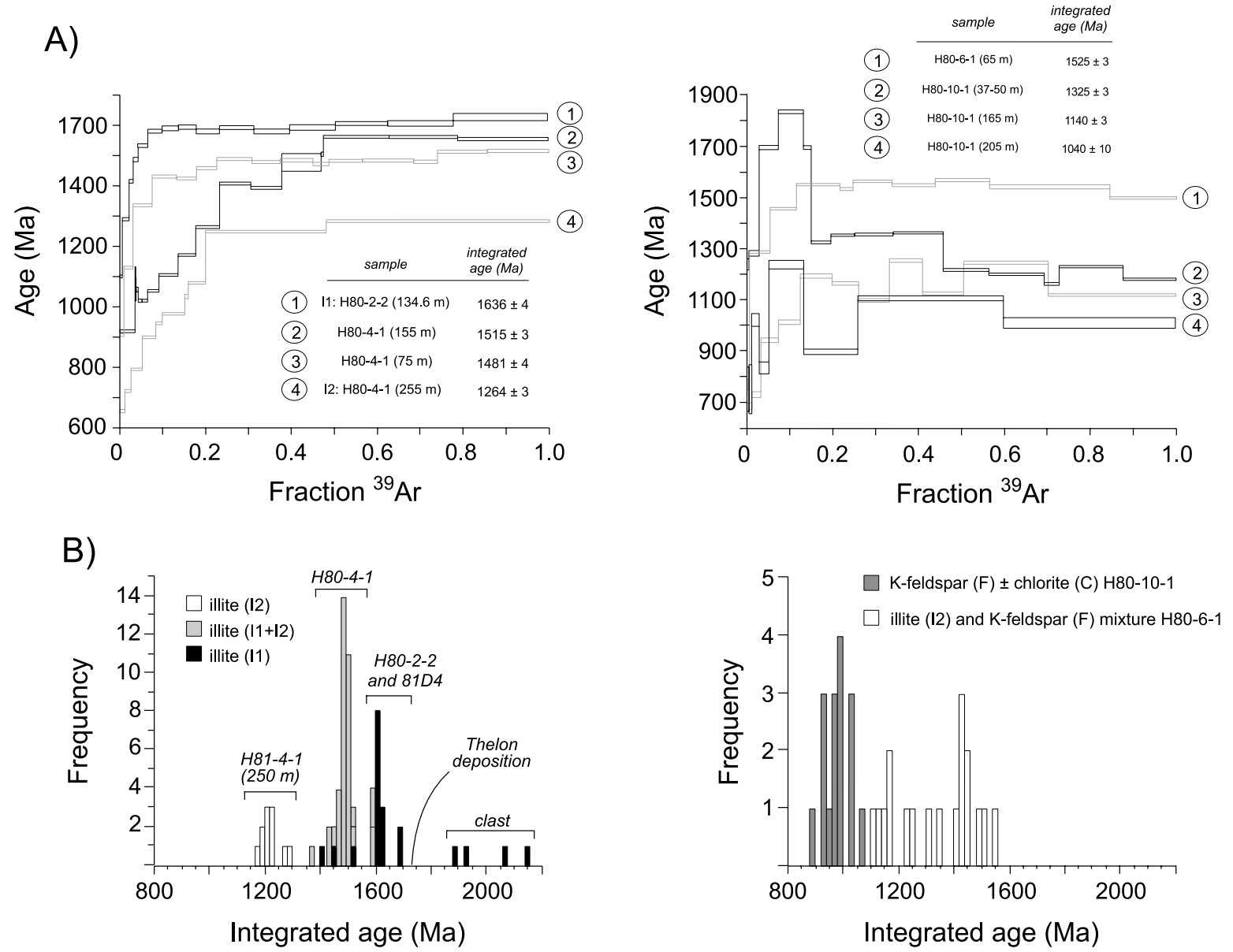


Table 1. Individual spot fusion ages ($\text{Ma} \pm 2\sigma$) from illite-rich (H80-6-1-62m, H80-2-2-136.4m, and H80-4-1-70m, H80-4-1-150m, H80-4-1-250m) and K-feldspar cemented sandstones (H80-10-1-39m, H80-10-1-163m and H80-10-1-205m), along with argon isotope ratios, % ^{40}Ar atmospheric contents and % ^{39}Ar released.

Sample	$^{36}\text{Ar}/^{40}\text{Ar}$	$^{39}\text{Ar}/^{40}\text{Ar}$	% $^{40}\text{Ar}_{\text{atm}}$	% ^{39}Ar	$^{40}\text{Ar}^*/^{39}\text{K}$	Age (Ma)
H80-6-1-62 m (F, I2)	Vol. $^{39}\text{Ar} = 48.51 \times 10^{-10}$			IA = 1444 ± 3 Ma		
Spot1	0.00007989	0.030422	2.35	0.97	32.096 ± 1.238	1171.6 ± 33.2
Spot2	0.00006875	0.02808	2.02	4.36	34.892 ± 0.408	1245.1 ± 10.5
Spot3	0.00014567	0.024458	4.29	22.06	39.132 ± 0.185	1351.2 ± 4.5
Spot4	0.00011969	0.023227	3.53	2.16	41.533 ± 1.106	1408.6 ± 26.0
Spot5	0.00010021	0.022935	2.95	18.78	42.313 ± 0.205	1426.9 ± 4.8
Spot6	0.00022828	0.021833	6.72	6.98	42.723 ± 0.238	1436.4 ± 5.5
Spot7	0.00008339	0.022695	2.45	2.71	42.981 ± 0.483	1442.4 ± 11.1
Spot8	0.00014509	0.021962	4.28	2.27	43.585 ± 0.844	1456.3 ± 19.3
Spot9	0.00015665	0.02103	4.6	7.98	45.363 ± 0.267	1496.6 ± 6.0
Spot10	0.00014309	0.020757	4.22	3.28	46.142 ± 0.580	1514.0 ± 12.9
Spot11	0.00010418	0.020479	3.07	28.44	47.331 ± 0.259	1540.2 ± 5.7
H80-10-1-39 m (F)	Vol. $^{39}\text{Ar} = 15.52 \times 10^{-10}$			IA = 1018 ± 5 Ma		
Spot1	0.00009026	0.038827	2.64	29.7	25.074 ± 0.160	966.1 ± 4.8
Spot2	0.00008022	0.038091	2.35	20.56	25.636 ± 0.354	982.7 ± 10.5
Spot3	0.0000637	0.03633	1.86	24.3	27.013 ± 0.226	1023.0 ± 6.5
Spot4	0.00012577	0.033986	3.67	15.62	28.344 ± 0.253	1061.0 ± 7.1
Spot5	0.00015524	0.030081	4.51	9.83	31.745 ± 1.025	1154.8 ± 27.5
H80-10-1-163 m (F > I2)	Vol. $^{39}\text{Ar} = 27.36 \times 10^{-10}$			IA = 1083 ± 4 Ma		
Spot1	0.00019643	0.041195	5.79	0.68	22.867 ± 6.001	899.9 ± 186.0
Spot2	0.00003613	0.039705	1.06	7.31	24.918 ± 0.613	962.4 ± 18.3
Spot3	0.00003173	0.03803	0.93	15.94	26.049 ± 0.210	995.9 ± 6.2
Spot4	0.00003833	0.036548	1.13	14.95	27.053 ± 0.214	1025.2 ± 6.2
Spot5	0.00008529	0.035975	2.5	3.42	27.101 ± 2.095	1026.6 ± 60.5
Spot6	0.00015667	0.031843	4.58	3.69	29.964 ± 1.219	1107.4 ± 33.7
Spot7	0.00002487	0.032519	0.73	50.19	30.526 ± 0.172	1122.9 ± 4.7
Spot8	0.00037156	0.023414	10.88	1.68	38.057 ± 1.077	1318.2 ± 26.46
Spot9	0.00025569	0.021702	7.47	2.15	42.634 ± 0.971	1427.3 ± 22.5
H80-10-1-205 m (F)	Vol. $^{39}\text{Ar} = 39.96 \times 10^{-10}$			IA = 994 ± 3 Ma		
Spot1	0.00005111	0.041202	1.5	18.66	23.906 ± 0.120	929.8 ± 3.6
Spot2	0.00006292	0.040575	1.85	4.09	24.190 ± 0.738	938.4 ± 22.3
Spot3	0.00011988	0.039836	3.52	7.1	24.220 ± 0.243	939.3 ± 7.4
Spot4	0.00005387	0.040408	1.58	8.24	24.356 ± 0.315	943.5 ± 9.5
Spot5	0.00002485	0.039363	0.73	4.94	25.219 ± 0.582	969.3 ± 17.3
Spot6	0.00003252	0.03867	0.95	24.89	25.613 ± 0.177	981.0 ± 5.2
Spot7	0.00008387	0.037167	0.73	4.94	25.219 ± 0.582	999.3 ± 3.8
Spot8	0.00012944	0.029578	3.78	6.03	32.529 ± 0.317	1174.5 ± 8.4
Spot9	0.00019756	0.027012	5.79	5.33	34.874 ± 0.299	1235.7 ± 7.7
H80-2-2-134.6 m (I1)	Vol. $^{39}\text{Ar} = 105.2 \times 10^{-10}$			IA = 1613 ± 3 Ma		
Spot1	0.00006503	0.01914	1.92	30.63	51.244 ± 0.290	1607.9 ± 6.0
Spot2	0.00005392	0.019201	1.59	14.16	51.251 ± 0.262	1608.0 ± 5.4
Spot3	0.00007067	0.019057	2.09	11.29	51.380 ± 0.284	1610.7 ± 5.9
Spot4	0.00005747	0.019116	1.7	11.62	51.423 ± 0.266	1611.6 ± 5.5
Spot5	0.00005744	0.01911	1.7	3.95	51.437 ± 0.254	1611.9 ± 5.3
Spot6	0.00004961	0.019156	1.46	8.05	51.438 ± 0.331	1611.9 ± 6.9
Spot7	0.00005412	0.019111	1.6	6.8	51.488 ± 0.222	1612.9 ± 4.6
Spot8	0.00005392	0.019201	1.63	8.15	51.855 ± 0.229	1620.5 ± 4.7
Spot9	0.00005643	0.018923	1.67	10.59	51.965 ± 0.364	1622.8 ± 7.5
Spot10	0.00005324	0.018919	1.57	4.76	52.025 ± 0.419	1624.0 ± 8.6
H80-4-1-70 m (I2)	Vol. $^{39}\text{Ar} = 39.83 \times 10^{-10}$			IA = 1485 ± 3 Ma		
Spot1	0.00006578	0.022428	1.93	11.05	43.726 ± 0.271	1454.3 ± 6.2
Spot2	0.00011624	0.021955	3.42	16.27	43.990 ± 0.235	1460.3 ± 5.3
Spot3	0.00005375	0.022355	1.58	13.21	44.026 ± 0.224	1461.1 ± 5.1
Spot4	0.00005417	0.022007	1.59	7.32	44.717 ± 0.269	1476.8 ± 6.1
Spot5	0.00006624	0.021714	1.94	7.00	45.160 ± 0.347	1486.7 ± 7.8
Spot6	0.0000705	0.021682	2.07	3.02	45.166 ± 0.610	1486.8 ± 13.7

Table 1. (concluded).

Sample	$^{36}\text{Ar}/^{40}\text{Ar}$	$^{39}\text{Ar}/^{40}\text{Ar}$	% $^{40}\text{Ar}_{\text{atm}}$	% ^{39}Ar	$^{40}\text{Ar}^*/^{39}\text{K}$	Age (Ma)
Spot7	0.00008951	0.021544	2.63	8.26	45.197 ± 0.263	1487.5 ± 5.9
Spot8	0.00006387	0.021421	1.87	19.69	45.810 ± 0.231	1501.2 ± 5.1
Spot9	0.00005275	0.021196	1.54	10.16	46.451 ± 0.281	1515.5 ± 6.2
Spot10	0.00005159	0.019773	1.5	4.02	49.814 ± 0.432	1588.2 ± 9.2
H80-4-1-150 m (I2)	Vol. $^{39}\text{Ar} = 68.27 \times 10^{-10}$		IA = 1504 ± 3 Ma			
Spot1	0.00005406	0.022117	1.58	1.19	44.500 ± 1.396	1475.1 ± 31.6
Spot2	0.00005845	0.021811	1.7	10.02	45.068 ± 0.205	1487.9 ± 4.6
Spot3	0.00009711	0.021549	2.85	3.4	45.083 ± 0.426	1488.2 ± 9.6
Spot4	0.00005062	0.021801	1.47	3.08	45.195 ± 0.520	1490.8 ± 11.7
Spot5	0.00003935	0.021764	1.15	4.5	45.420 ± 0.292	1495.8 ± 6.5
Spot6	0.00006684	0.021409	1.95	18.16	45.800 ± 0.212	1504.3 ± 4.7
Spot7	0.0000734	0.021801	2.13	5.78	46.075 ± 0.274	1510.4 ± 6.1
Spot8	0.00006799	0.021153	1.98	11.09	46.335 ± 0.291	1516.2 ± 6.4
Spot9	0.0001118	0.02071	3.26	40.17	46.711 ± 0.220	1524.5 ± 4.9
Spot10	0.00010914	0.020458	3.2	1.1	47.315 ± 0.967	1537.7 ± 21.1
Spot11	0.00034719	0.018118	10.2	1.52	49.562 ± 0.574	1586.2 ± 12.2
H80-4-1-250 m (I2)	Vol. $^{39}\text{Ar} = 33.73 \times 10^{-10}$		IA = 1217 ± 3 Ma			
Spot1	0.00004723071	0.030526474	1.39	6.86	32.274 ± 0.139	1175.8 ± 3.7
Spot2	0.00004518825	0.030106084	1.33	7.92	32.745 ± 0.238	1188.4 ± 6.3
Spot3	0.0000406973	0.029943534	1.199	11.34	32.967 ± 0.164	1194.2 ± 4.4
Spot4	0.0000444766454	0.029558977	1.308	11.35	33.360 ± 0.101	1204.6 ± 2.7
Spot5	0.00008814006	0.028960385	2.587	1.55	33.609 ± 0.932	1211.2 ± 24.5
Spot6	0.00004189729	0.029296965	1.232	26.63	33.684 ± 0.113	1213.1 ± 2.9
Spot7	0.00005884467188	0.028721463	1.728	17.95	34.187 ± 0.084	1226.3 ± 2.2
Spot8	0.00004739467	0.028704382	1.395	4.00	34.323 ± 0.431	1229.8 ± 11.2
Spot9	0.0000893	0.02899687	2.63	5.41	34.487 ± 0.398	1234.1 ± 10.3
Spot10	0.0000357428	0.027523216	1.05	2.49	35.922 ± 0.356	1270.9 ± 9.1
Spot11	0.00003373528	0.02725506	0.99	4.5	36.298 ± 0.454	1280.4 ± 11.4
81-D-4-96 m Clast	Vol. $^{39}\text{Ar} = 16.60 \times 10^{-10}$		IA = 2098 ± 12 Ma			
Spot1	0.000041776	0.016809	1.23	8.91	58.758 ± 0.963	1933.9 ± 19.4
Spot2	0.00007562	0.014756	2.23	47.33	66.255 ± 0.728	2079.2 ± 13.6
Spot3	0.000302	0.013001	8.92	43.76	70.056 ± 0.939	2148.7 ± 16.7
81-D-4-225 m (I1)	Vol. $^{39}\text{Ar} = 4.71 \times 10^{-10}$		IA = 1645 ± 10 Ma			
Spot1	0.00020757	0.023301	6.13	9.88	40.286 ± 2.273	1516.1 ± 57.8
Spot2	0.00008339	0.022131	2.46	49.34	44.072 ± 0.674	1610.0 ± 16.3
Spot3	0.00008187	0.020645	2.42	38.37	47.267 ± 0.794	1685.6 ± 18.4
Spot4	0.00019318	0.019859	5.7	2.31	47.482 ± 7.204	1690.6 ± 166.5
81-D-4-377.3 m Vein	Vol. $^{39}\text{Ar} = 18.61 \times 10^{-10}$		IA = 1449 ± 10 Ma			
0.25W	0.00013222	0.027546	3.9	7.21	34.885 ± 0.828	1373.3 ± 22.8
0.50W	0.00002538	0.025933	0.75	66.4	38.272 ± 0.481	1464.3 ± 12.6
1.50W	0.00006381	0.027317	1.88	0.57	35.917 ± 6.479	1401.5 ± 175.8
7.00W (TF)	0.00035117	0.015835	10.37	0.21	56.602 ± 18.529	1890.2 ± 383.4

Note: Total ages of gas extracted are presented as IA (integrated ages) in Ma. Vol., volume in cm^3 .

$130 \pm 30^\circ\text{C}$ and final ice melting temperatures ca. $-17 \pm 2^\circ\text{C}$ (Table 2). All the inclusions in the overgrowths homogenized from liquid and vapor to liquid, with no consistent variation in homogenization or ice melting temperatures over the 300-m range in depth. Final ice melting temperatures were only a few degrees above initial ice melting, implying that NaCl was the major dissolved constituent in the fluid. The final ice melting temperatures correspond to fluid salinities of ca. 17 wt.% equivalent NaCl (Fig. 7A). Assuming an average geothermal gradient of ca. $40^\circ\text{C}/\text{km}$ for the Thelon Basin, quartz overgrowth precipitation at ca. 130°C corresponds to a depth of ca. 2.5 km.

Quartz veins are rare, as are primary fluid inclusions within

them, and are restricted to the silicified, basal units of the Thelon Formation. Although the data are limited, primary fluid inclusions in Q2 veins from the edge of the basin are 2-phase (liquid and vapor), have homogenization temperatures from 113°C to 231°C , and initial and final ice-melting temperatures of -20°C to -11°C , similar to the properties of the inclusions in the overgrowths (Table 2). Inasmuch as euhedral quartz veins postdate the overgrowths but predate peak-diagenetic illite, the veins probably represent activation of structures in the basin prior to maximum burial of the basal units.

Although K-feldspar contains some fluid inclusions along cleavage planes, none of these were appropriate for ther-

Table 2. Temperatures of homogenization (Th), initial temperatures of ice melting (Tfm), and average salinities obtained from final ice melting (Tice) for primary fluid inclusions in quartz overgrowths (Q1), quartz veins (Q2), vein filling dolomite (D), and for secondary inclusions in quartz from the cores of rhombic K-feldspar (F).

Drillhole	Formation	depth (m)	Mineral and petrography	Th (°C) mean $\pm$ 2 σ (no. of samples)	Tfm (°C) mean $\pm$ 2 σ (no. of samples)	Tice (°C) mean $\pm$ 2 σ (no. of samples)	Average salinity (wt.% NaCl equivalent)
H80-3-1	Pre-Thelon	35.0	Quartz vein	113 $\pm$ 12 (5)	n.d.	-14.5 (1)	18.2
			Late dolomite (vein filling)	<50 (10)	n.d.	n.d.	
H80-2-2	Thelon	26.0	Quartz vein	132 $\pm$ 20 (22)	n.d.	-20.8 $\pm$ 2 (5)	18.6
			Late dolomite (vein filling)	98 $\pm$ 15 (6)	n.d.	-15.3 $\pm$ 1 (4)	18.9
H80-4-1	Thelon	150.0	Q1	159 $\pm$ 36 (41)	n.d.	-11.4 $\pm$ 3 (5)	15.4
		250.0	Q2	231 $\pm$ 17 (6)	n.d.	n.d.	
H80-6-1	Thelon	62.0	Q1	115 $\pm$ 23 (10)	n.d.	-16.4 $\pm$ 5 (5)	19.7
			Dolomite cement (D)	103 $\pm$ 36 (7)	n.d.	-2.9 $\pm$ 1 (5)	4.7
H80-10-1	Thelon	163.0	Q1	184 $\pm$ 32 (24)	-23.3 $\pm$ 3 (2)	-16.2 $\pm$ 4 (6)	19.5
			F	99 $\pm$ 26 (6)	n.d.	-18.8 $\pm$ 1 (2)	21.5

Note: n.d., not determined.

mometry. However, SEM and petrographic examination of K-feldspar revealed that some contained residual quartz in their cores with secondary fluid inclusions aligned with those in the cleavage planes of the host feldspar. These secondary inclusions in the quartz are 2-phase (liquid and vapor), have homogenization temperatures of between 60°C and 110°C, and have final ice melting temperatures of -19°C, which correspond to salinities of 21.5 wt.% equivalent NaCl (Table 2; Fig. 7A). Assuming that the secondary inclusions in the quartz are equivalent to primary fluid inclusions in the feldspar, the feldspar would have formed from a brine with the same salinity as that for the overgrowths and quartz veins, but at lower temperatures. These lower temperatures of formation for the feldspar are consistent with the temperatures deduced from their crystal structure. In addition, the salinities of the fluids are similar to those reported by Girard et al. (1988) for authigenic feldspar in recent basins.

Homogenization temperatures of 2-phase, primary fluid inclusions in dolomite (D) range from 90°C to 120°C. Final ice melting temperatures range from -4°C to -1°C, corresponding to salinities between 1 and 5 wt.% equivalent NaCl (Table 2; Fig. 7A). The distinctly lower salinities for the late dolomite fluid relative to any before indicate that there was incursion of meteoric waters late in the evolution of the basin.

Stable isotope results

Detrital quartz (Q0) from the Thelon Formation has $\delta^{18}\text{O}$ values from 9.5‰ to 10.6‰ (Table 3), whereas detrital quartz from the underlying pre-Thelon Formation Sandstone range from 11‰ to 14‰. Differences in $\delta^{18}\text{O}$ values imply different provenance for these formations.

Kaolinite (K), which occurs in veins in the regolith, has $\delta^{18}\text{O}$ values near 11‰ and δD values of ca. -45‰ (Table 3). No pure separates could be obtained, so these values represent those calculated for end-member kaolinite from mixtures of illite and kaolinite. Although this occurrence of kaolinite is petrographically consistent with formation at low temperatures during weathering of the regolith, the $\delta^{18}\text{O}$ values are too low, or the δD values too high, for those of pedogenic kaolinites (Lawrence and Taylor 1972).

During the earliest stage of diagenesis in the Thelon Basin, hematite (H1) and then quartz overgrowths (Q1) were precipitated. Two samples of specular hematite (Table 3) have $\delta^{18}\text{O}$ values of -3.6‰. These isotopic compositions are similar to those that produced the earliest diagenetic hematite in the Athabasca Basin between 1600 and 1750 Ma (Kotzer et al. 1992). Quartz overgrowths from the Thelon Formation have $\delta^{18}\text{O}$ values between 11‰ and 14‰, also identical to those from the Athabasca Basin (Kotzer and Kyser 1995). If the hematite and the quartz overgrowths formed from the same fluid, the difference in their isotopic compositions would require unrealistic temperatures in excess of 300°C. Using temperatures of ca. 130°C for the fluids from which the quartz overgrowths in the Thelon Basin precipitated (Table 2), the calculated $\delta^{18}\text{O}$ values for this fluid is from -3‰ to +1‰ (Table 3). The $\delta^{18}\text{O}$ values of the fluids that would have formed the hematite at temperatures of ca. 50–100°C would have been much lower (Table 3).

Quartz veins (Q2) from the basal Thelon and pre-Thelon Formation Sandstone (Table 3) have $\delta^{18}\text{O}$ values of 12‰ in

Fig. 7. Comparison between the (A) temperature and salinity and (B) δD and $\delta^{18}O$ values of fluids associated with specific diagenetic minerals in the Thelon and Athabasca Basins. Q, quartz; F, feldspar; S, siderite; C, chlorite; I, illite; D, dickite; U, uraninite; K, kaolinite, and their numbers indicate the order of appearance for each type of mineral. Compositions of Thelon fluids are shaded areas, whereas those for the Athabasca are not. Also shown are the inferred ages of the fluids, the evolution of the composition of fluids in each basin (Thelon is solid line, Athabasca dashed line), and the isotopic compositions of global meteoric waters (meteoric water line (MWL)), modern groundwaters in the basins, and ocean waters (standard mean ocean water (SMOW)). Data for the Thelon Basin are from this study (Tables 2 and 3), and those for the Athabasca are primarily from Kotzer and Kyser (1995).

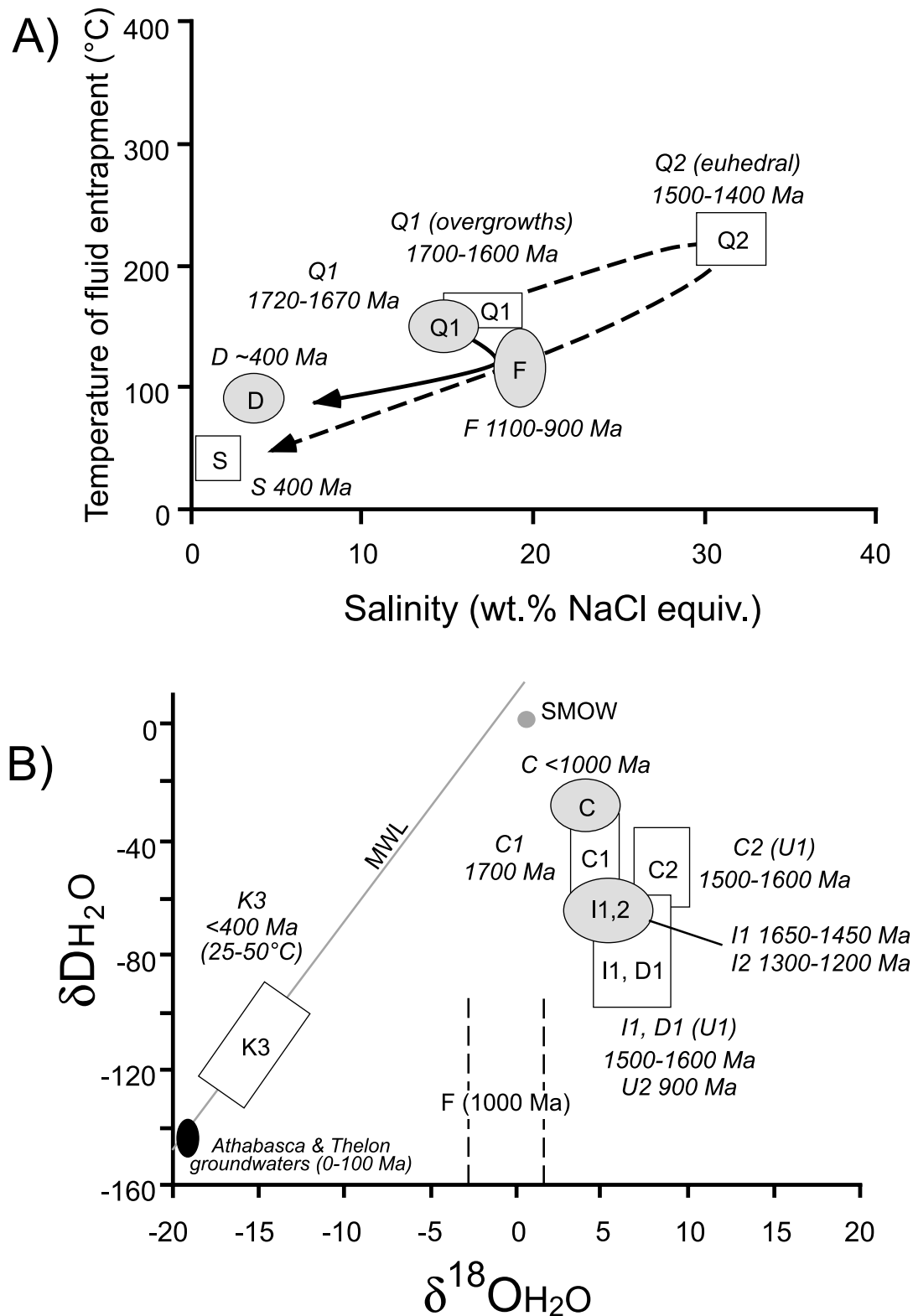


Table 3. Isotopic compositions (δD , $\delta^{18}O$, and $\delta^{13}C$ values) of detrital quartz (Q0), hematite (H1), vein kaolinite (K) in the regolith, quartz overgrowths (Q1), vein quartz (Q2), illite (I1, I2), K-feldspar (K), chlorite (C), and vein dolomite (D) from the Thelon sub-basin. The δD and $\delta^{18}O$ values of fluids in equilibrium with these minerals were calculated assuming temperatures of formation of 100°C for Q1 and H1, 200°C for illite, 100°C for K-feldspar, 75°C for Al-Mg chlorite, and 100°C for dolomite.

Drill hole	Depth (m)	Mineral	Description	δD min	% H ₂ O	$\delta^{13}C$	$\delta^{18}O$ min	δD water	$\delta^{18}O$ water
81-D-2	29.5	I1	<2 μm	-65	5.5		15.1	-48	9.0
	58.2	90 (I1)/10(K)	<2 μm	-61	5.8		16.1	-45	10.0
	91.4	I1	<2 μm	-62	5.2		15.2	-45	9.1
	127.8	I1	<2 μm	-64	5.5		15.7	-47	9.6
	137.5	H2	Specular hematite				8.2		-2.8
		Q1	Overgrowth				16.4		-2.0
		I1	<2 μm	-62	5.1		14.7	-45	8.6
81-D-4	43.7	I1	<2 μm	-48	6.1		16.2	-31	10.1
	96.0	Q1	Overgrowth				16.6		-1.8
		I1	<2 μm	-64	6.1		15.4	-47	9.3
	225.0	I1	<2 μm	-60	4.1		14.6	-44	8.5
	260.0	60 (I1)/40 (K1)	<2 μm	-54	6.9		13.2		
	266.0	60 (I1)/40 (K1)	<2 μm	-50	7.0		14.4		
	322.8	95 (I1)/10 (K1)	<2 μm	-57	5.1		14.8	-41	8.7
H80-2-2	26.0	Q0	Detrital quartz				11.2		
		Q1	Quartz vein				14.5		0.0
		I1	<2 μm	-76	5.5		13.2	-60	7.1
		D	Dolomite vein			-0.5	17.2		-6.0
	60.0	D	Dolomite vein			-1.9	16.7		-6.5
	72.5	I1	<2 μm	-76	5.2		13.0	-60	6.9
	94.0	I1	<2 μm	-71	5.4		12.4	-54	6.4
	136.4	Q0	Detrital quartz				13.3		
		I1	<2 μm	-70	5.3		11.6	-54	5.5
		I1	<2 μm (mud clast)	-71	5.4		11.7	-54	5.6
H80-4-1	70.0	I1	<2 μm	-61	5.5		12.8	-45	6.7
	90.0	I1	<2 μm	-70	5.6		13.2	-54	7.2
	110.0	H1	Specular hematite				-3.6		-11.6
		I1	<2 μm	-63	4.8		11.3	-47	5.2
	150.0	Q1	Quartz overgrowth				13.8		-0.7
		90 (I1)/IS(10)	<2 μm	-60	8.8		10.0	-44	3.9
	210.0	Q1	Quartz overgrowth				13.9		-1.0
	250.0	Q0	Detrital quartz				10.1		
		Q2	Quartz overgrowth				12.0		-2.0
		I2	<2 μm	-73	6.8		13.4	-57	7.3
H80-6-1	22.0	I1	<2 μm	-72	5.0		14.2	-56	8.1
	42.0	I1	<2 μm	-62	5.7		13.8	-46	7.7
	62.0	Q0	Detrital quartz				10.8		
		Q1	Euhedral quartz				16.1		1.0
		70 (I1)/ 30 (C)	<2 μm	-66	6.9		15.0		
		70 (F)/ 30 (Q0)	500–250 μm				14.6		-3.0
		D	Dolomite cement			-2.9	17.4		-6.0
	94.7	I1	<2 μm	-78	5.6		15.0	-62	8.9
	131.0	65 (I1)/ 25 (C)	<2 μm	-63	6.3		13.2		
	172.0	Q1	Quartz overgrowth				11.3		-3.7
H80-10-1		95 (I1)/ 5 (C)	<2 μm	-66	5.7		14.9	-50	8.8
	202.0	70 (I1)/30 (C)	<2 μm	-65	7.0		14.3		
	207.0	60 (I1)/ 35 (C)	<2 μm	-58	7.8		13.5		
	226.0	70 (I1)/ 30 (C)	<2 μm	-64	7.2		13.2		
	31.0	H1	Specular hematite				-3.6		-11.6
		Q1	Euhedral quartz				13.5		-0.5
		90 (C)/10 (I1)	<2 μm	-51	11.8		12.0	-23	3.1
	34.0	80 (C)/ 20 (I1)	<2 μm	-50	11.0		11.8	-22	2.9
	39.0	Q0	Detrital quartz				10.0		
		Q1	Quartz overgrowth				11.3		-2.7

Table 3. (concluded).

Drill hole	Depth (m)	Mineral	Description	δD min	% H ₂ O	$\delta^{13}C$	$\delta^{18}O$ min	δD water	$\delta^{18}O$ water
		70 (C)/30 (I1)	<2 μm	-69	10.6		10.7		
	93.0	65 (C)/ 35 (I1)	<2 μm	-48	10.6		11.6		
	114.0	80 (C)/20 (I1)	<2 μm	-52	11.6		11.3	-24	2.4
	135.0	Q1	Euhedral quartz				15.3		-0.7
	163.0	Q1	Quartz overgrowth				11.3		-2.7
		70 (C)/30 (I)	<2 μm	-54	10.6		11.9		
	205.0	Q0	Detrital quartz				10.4		
		Q1	Quartz overgrowth				12.6		-1.4
		80 (C)/20 (I)	<2 μm	-55	11.3		12.0	-27	3.1
		D	Dolomite cement			-2.7	16.0		-7.4

Note: min, minimum.

quartz having fluid inclusion homogenization temperatures of 130°C. The calculated $\delta^{18}O$ values of the fluids in equilibrium with these is ca. -2‰, similar to those that formed the earlier quartz overgrowths.

Temperatures for the formation of illite (I1 and I2) were estimated using their polytypes, as described above. The 2M₁ illites from the basal units are assumed to have formed at ca. 200°C. Although the later illites may have equilibrated at slightly lower temperatures because of the presence of some 1M polytype, a temperature of 200°C will be assumed because the lack of constraint on exactly how much lower the temperature would have been. The $\delta^{18}O$ values for illite from the basal units (I1) vary from 14.6‰ to 15.7‰ and their δD values from -71‰ to -48‰ (Table 3). In contrast, the less crystalline and later illite (I2) has slightly lower $\delta^{18}O$ values from 10.0‰ to 13.4‰ and δD values from -78‰ to -60‰. Isotopic compositions of peak-diagenetic fluids in equilibrium with the earliest illite (I1) were between 8.5‰ to 9.6‰ in $\delta^{18}O$ and -54‰ to -31‰ in δD at 200°C. Fluids in apparent isotopic equilibrium with the later illites (I2) would have had $\delta^{18}O$ values from 3.9‰ to 7.3‰ and δD values from -62‰ to -44‰. If the temperatures of formation of the I2 illites were lower by ca. 50°C, the calculated isotopic compositions of fluids from which they formed would be identical to those for the I1 illites. The high $\delta^{18}O$ values of these fluids are typical of basinal brines that have had extensive water-rock interaction (e.g., Kyser and Kerrich 1990). Their δD values, which are normally unaffected by water-rock interactions, are consistent with an origin for the fluids as low-latitude meteoric waters. Both the isotopic compositions and origin of the peak-diagenetic fluids in the Thelon Basin are similar to those in the Athabasca Basin (Kotzer and Kyser 1995; Kyser et al. 2000).

Feldspar (F) from the Thelon Formation has a calculated $\delta^{18}O$ value of 16‰ (Table 3). Using the temperature of 60°C to 110°C obtained from the fluid inclusions, the value calculated for the fluid would be between -3‰ and 1‰ (Table 3; Fig. 7B). These fluids have oxygen isotopic compositions that differ markedly from those associated with peak diagenesis and indicate a different origin for the fluids. The Al-Mg chlorite (C) has calculated $\delta^{18}O = 15 \pm 0.5$ ‰ and $\delta D = -58 \pm 10$ ‰. The polytype of the chlorite and the lack of reaction between the preexisting feldspar and the fluid that precipitated the chlorite constrain the temperature to ca. 75°C for formation of the sudoite. The chlorite would have formed from fluids having $\delta^{18}O$ values near 5‰ and δD near -30 ‰ (Fig. 7B).

These are isotopically and chemically distinct from the fluids that equilibrated with the feldspar.

Dolomites from veins have $\delta^{13}C$ values from 0‰ to -3‰, which indicate a predominantly inorganic source of carbon. The $\delta^{18}O$ values of the dolomites range from 16.7‰ to 19.3‰. Using fluid inclusion homogenization temperatures of 100°C in the dolomite veins, the $\delta^{18}O$ values of the fluid in equilibrium with the dolomite in the veins would have been -8‰ to -4.5‰. These fluids are the last recorded by the diagenetic minerals in the Thelon Basin, and also have the lowest salinities (Fig. 7A) and $\delta^{18}O$ values.

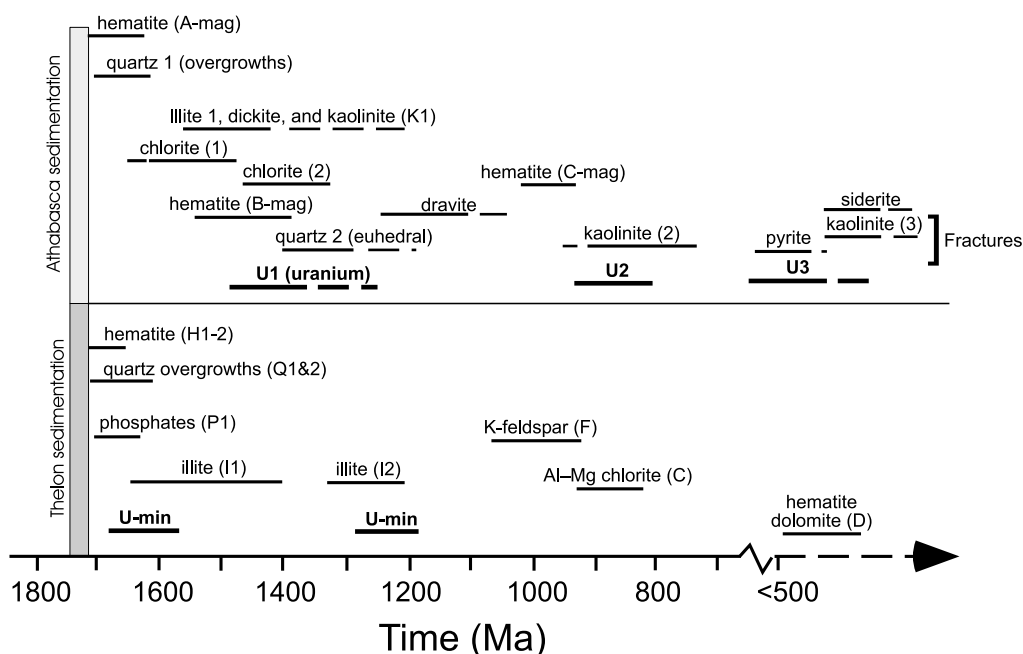
Discussion

The Thelon Basin has a protracted fluid history recorded in the distinct diagenetic phases throughout the basin from which information about the characteristics of diagenetic fluids have been extracted (Fig. 7). Sedimentation in the basin started prior to 1720 Ma, which is the age of early authigenic phosphate sporadically distributed throughout the basin, and after 1753 Ma, the youngest age for last magmatic event in the basement (Miller et al. 1989). Basin sedimentation, therefore, began during the Paleoproterozoic, although peak diagenesis occurred dominantly during the Mesoproterozoic at 1400–1650 Ma.

Early kaolinite is preserved in veins in the accreted regolith between the Archean–Paleoproterozoic basement rocks and the overlying clastic sediments of the Thelon Formation. From geologic and petrographic relations, this kaolinite was originally a weathering product of the regolith, but isotopic compositions of the kaolinite are not consistent with this, perhaps because of interaction with later fluids. Early diagenesis resulted in precipitation of extensive hematite on the surfaces of detrital quartz grains throughout the Thelon Formation. Diagenesis then resulted in syntaxial quartz (Q1) on detrital quartz (Q0) and detrital feldspar, and later quartz veining (Q2), at 100–160°C from NaCl brines having ca. 17 wt.% equivalent NaCl (Table 2). The $\delta^{18}O$ value of this brine was ca. 0‰, and it was distributed throughout the Thelon sub-basin. Silicification of this type is most pronounced in basal and fine-grained sequences throughout the Thelon Basin, and these may have acted as regional aquitards during later fluid circulation (Hiatt et al. 1999).

A period of extensive desilicification followed, during which the remaining detrital feldspars and many of the quartz overgrowths were dissolved. Dissolution most likely occurred

Fig. 8. Comparison between the mineral, temporal, and thermal evolution during diagenesis of the Thelon and Athabasca Basins. Both basins have had a protracted history of diagenesis that was similar initially and then diverged near peak diagenesis. Data for the Thelon Basin are from this study, and those for the Athabasca are primarily from Kotzer and Kyser (1995).



in response to an increase in temperature of the pore fluids as burial proceeded. Peak diagenesis began at ca. 1690 Ma and is manifest by the formation of illite (I1) at ca. 200°C preserved in the basal units of the Thelon Formation. Peak-diagenetic fluids were active in some areas of higher stratigraphic units until ca. 1150 Ma, the youngest Ar–Ar ages measured in the diagenetic illite (I2). There was either an evolution in the $\delta^{18}\text{O}$ values of these fluids, from ca. 9‰ for illite (I1) in the pre-Thelon Formation Sandstone and basal Thelon Formation to ca. 3.9‰ for illite (I2) in the more permeable upper stratigraphic units, or a decrease in temperature of ca. 50°C (Figs. 7A, 7B). However, δD values of the fluids were relatively constant at ca. $-50 \pm 15\text{‰}$, which are values not unlike those of low-latitude meteoric waters. The substantially higher $\delta^{18}\text{O}$ values of the peak-diagenetic fluids relative to those for the quartz overgrowths imply that the peak-diagenetic fluids represent ^{18}O -enrichment as a result of more extensive water–rock interaction, most likely during evolution of the basin.

Late diagenesis in the Thelon Basin is reflected by authigenic K-feldspar (F) and then sudoite (C), which crystallized from saline brines at ca. 1000 Ma and 100°C in the central portion of the sub-basin. Fluids related to the K-feldspar had high salinities of 22 wt.% equivalent NaCl and $\delta^{18}\text{O}$ values of -3‰ to 1‰ whereas those for the chlorite had much higher $\delta^{18}\text{O}$ values ca. 5‰ and δD values of -30‰ (Fig. 7B). The mineralogy, high salinities and heavy isotopic compositions of these authigenic minerals are consistent with an evaporitic source for the fluids, perhaps related to evaporitic sequences preserved in the western part of the basin.

The youngest fluid event preserved in the Thelon Basin is that associated with the precipitation of minor dolomite (D) in reactivated quartz veins (Q2) in the basal units. Although the exact age is uncertain, the fluid was ca. 100°C, with low

salinities between 1 and 5 wt.% equivalent NaCl, and low $\delta^{18}\text{O}$ values of -8‰ to -4.5‰ (Table 3). The distinct occurrence, salinity, and composition of this fluid imply that enough time had passed to totally expel the brines that had formed the chlorite. The dolomite represents the last distinct fluid event preserved in the authigenic minerals of the Thelon Basin.

Comparison with the contemporaneous Athabasca Basin

The Athabasca Basin hosts world-class unconformity-type uranium deposits that formed as a result of extended diagenetic and hydrothermal fluid histories (Kotzer and Kyser 1995). It is spatially and temporally related to the Thelon Basin, but large uranium deposits in the Thelon Basin have yet to be discovered. Despite similarities in the sedimentological and evolutionary record of these two basins (e.g., Miller 1995), they have some differences in their diagenetic histories that may have implications for their economic mineral potential in relation to the evolution of ancient intracratonic basins.

Although the timing of sedimentation in the Athabasca Basin is less certain than in the Thelon Basin, early diagenesis in both basins is virtually the same (Fig. 8). Both begin with detrital quartz coating by early diagenetic hematite (H). This hematite in the sandstones from the Athabasca has a paleomagnetic direction consistent with that of the earth at 1600–1750 Ma. Declinations similar to these have been found in sandstones from drill core in the Thelon Basin (Dobrowsky et al. 1994), and these may reflect the same general time of early diagenetic hematite formation in both basins. In addition, $\delta^{18}\text{O}$ values of ca. -3‰ for the early hematite in the Athabasca Basin (Kotzer et al. 1992) are identical to those for the Thelon Basin, implying that both

the timing and isotopic composition of these fluids were similar. The hematite in both basins is followed by an extensive period of quartz overgrowths (Q1) on detrital quartz. In the Thelon Basin, there is evidence for detrital feldspar grains at this time, but there is no evidence that they were present in the Athabasca sediments. The fluids associated with the overgrowths in both basins had similar salinities, temperatures (Fig. 7), and $\delta^{18}\text{O}$ values.

It is during peak diagenesis that the evolutions of the Thelon and Athabasca Basins start to become distinct. Although illite is a common peak-diagenetic mineral in the both basins and seems to have formed from fluids having similar properties and isotopic compositions, peak-diagenetic assemblages are more complex in the Athabasca Basin (Fig. 8). The peak-diagenetic assemblage in the Athabasca Formation is dickite and illite, with minor dravite, hematite, goyasite, and clinocllore, rather than simply illite. In addition, peak diagenesis in the Athabasca Basin is synchronous with formation of world-class unconformity-type uranium deposits and extensive quartz veins (Q2) that formed from fluids having high temperatures (ca. 250°C) and salinities (ca. 30 wt.% equivalent NaCl). Evidence for these types of fluids in the Thelon is either not preserved, not yet discovered, or the fluids did not exist. The fluids were different in composition for this period of time for these two basins (Fig. 7).

K-feldspar followed by chlorite precipitation represent late events at ca. 1000 Ma in the late diagenetic evolution of the Thelon Basin, but these are conspicuously absent in the Athabasca Basin (Fig. 8). The fluids responsible for these precipitates imply large-scale circulation in the central portion of the sub-basin of silica- and illite-undersaturated fluids that initially produced secondary porosity. Moreover, these fluids are chemically and isotopically distinct from any of the late diagenetic fluids that affected the Athabasca Basin (Fig. 7). Contemporary late diagenesis in the Athabasca Basin is manifested primarily as uranium mobilization and hematite recrystallization at ca. 900 Ma (U2; Fig. 8). The fluid associated with this event had similar salinities, temperatures, and isotopic compositions as that at peak diagenesis (Kotzer and Kyser 1995). Fluid movement in the basins around ca. 900–1000 Ma may have been facilitated by the break-up of Rodinia (Hoffman 1991; Ross et al. 1992; Blewett et al. 1998 and Idnurm and Giddings 1998).

Later influx along reactivated structures in both basins resulted in precipitation of vein-filling carbonates from fluids having relatively low salinities and temperatures (Fig. 8). These fluids were probably meteoric in origin based on their isotopic compositions. Incursion of meteoric waters into the Athabasca Basin primarily along reactivated structures resulted in alteration and precipitation of low-temperature kaolinite (Fig. 8). The isotopic composition of the fluids that produced these kaolinites are representative of high-latitude meteoric waters (Fig. 7; Kotzer and Kyser 1995). Finally, recent (i.e., <60 Ma) incursion of meteoric waters having low δD and $\delta^{18}\text{O}$ values along reactivated structures in the Athabasca Basin has resulted in the exchange of hydrogen isotopes between this fluid and hydrous minerals near fractures throughout the basin (Wilson and Kyser 1987; Kotzer and Kyser 1995), and total exchange of oxygen between the groundwaters and uraninite (Fayek and Kyser 1997 and 1999). No evidence of

these latter processes has been found in the sediments of the Thelon Formation.

Conclusions

The Thelon sub-basin is analogous to the Athabasca Basin with respect to its early diagenetic history (ca. 1750–1650 Ma); however it evolved in a geochemically distinct fashion as it reached peak diagenesis at ca. 1500 Ma. The peak-diagenetic assemblages in both basins were distinct, as were the maximum temperatures reached. The Thelon Formation has illite as the major diagenetic component, whereas the Athabasca has illite and dickite along with minor dravite, hematite, and phosphates. These results imply that the fluids in the Athabasca Basin had a different composition, particularly for K and some minor elements. The origin of these differences cannot be definitively ascertained, but could have resulted if the detritus contributed to the Athabasca was more intensely weathered (i.e., no detrital feldspar), the Athabasca was more deeply buried, or the original sedimentology of the Athabasca facilitated the formation of aquifers at depth.

Late diagenesis in both basins is markedly different. Whereas the peak-diagenetic fluid may have been resident in the Athabasca until ca. 900 Ma, when uranium was mobilized, there was influx of new brines into the Thelon Formation that generated secondary porosity and precipitated feldspar and then sudoite. These fluid events are temporally coeval and may have been initiated tectonically by the assemblage of Rodinia. Later events are manifest primarily in the Athabasca Basin as formation of kaolinite from meteoric waters, low-temperature exchange of hydrogen isotopes between water and hydrous minerals near fractures, and resetting of oxygen isotopes in uraninite. Although temporally and spatially related, these two Proterozoic basins have diagenetic and fluid evolutions that began to diverge during a critical interval of time.

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References

- Becker, R.H., and Clayton, R.N. 1976. Oxygen isotope study of a Precambrian banded iron-formation, Hamersley Range, Western Australia. *Geochimica et Cosmochimica Acta*, **40**: 1153–1165.
- Blewett R.S., Black, L.P., Sun, S.-S., Knutson, J., Hutton, L.J., and Bain, J.H.C. 1998. U–Pb zircon and Sm–Nd geochronology of the Mesoproterozoic of north Queensland: Implications for a

- Rodinia connection with the belt supergroup of North America. *Precambrian Research*, **89**: 101–127.
- Clayton, R., and Mayeda, T.K. 1963. The use of bromine pentafluoride in the extraction of oxygen from oxides and silicates for isotopic analysis. *Geochimica et Cosmochimica Acta*, **27**: 43–52.
- Dobrowsky, J.C., Kyser, T.K., and Baker, J. 1994. Paleomagnetism and fluid movement in the Athabasca Basin. Saskatchewan Geological Survey, 94-2.
- Donaldson, J.A. 1965. The Dubawnt Group, District of Keewatin and Mackenzie. Geological Survey of Canada, Paper 64-20.
- Donaldson, J.A. 1969. Descriptive notes (with particular reference to the late Proterozoic Dubawnt Group) to accompany a geological map of central Thelon Plain, District of Keewatin and Mackenzie. Geological Survey of Canada, Paper 68-49.
- Fayek, M., and Kyser, T.K. 1997. Characterization of multiple fluid-flow events and rare-earth-element mobility associated with formation of unconformity-type uranium deposits in the Athabasca Basin. *Canadian Mineralogist*, **35**: 627–658.
- Fayek, M., and Kyser, T.K. 1999. Stable isotope geochemistry of uranium deposits. In *Uranium: Mineralogy, geochemistry and the environment*. Edited by P.C. Burns and R. Finch. Reviews in Mineralogy, **38**: 181–220.
- Fuchs, H.D., and Hilger, W. 1989. Kiggavik (Lone Gull): an unconformity related uranium deposit in the Thelon Basin, Northwest Territories, Canada. In *Uranium resources and geology of North America*. Edited by E. Muller-Kahle. International Atomic Energy Agency (IAEA)-TECDOC-500, Vienna, pp. 429–454.
- Gall, Q. 1994. The Proterozoic Thelon paleosol, NWT, Canada. *Precambrian Research*, **68**: 115–137.
- Gall, Q., Peterson, T.D., and Donaldson, J.A. 1992. A proposed revision of Early Proterozoic stratigraphy of the Thelon and Baker Lake basins, Northwest Territories. In *Current research, part C*. Geological Survey of Canada, Paper 92-1 C, pp. 129–137.
- Gandhi, S.S. 1989. Geology and uranium potential of the Thelon Basin and adjacent basement in comparison with the Athabasca Basin region. In *Uranium resources and geology of North America*. Edited by E. Muller-Kahle. IAEA-TECDOC-500, Vienna, pp. 411–428.
- Girard, J.-P., Aronson, J.L., and Savin, S.M. 1988. Separation K-Ar dating and $^{18}\text{O}/^{16}\text{O}$ ratio measurements of diagenetic K-feldspar overgrowths: An example from the Lower Cretaceous arkoses of the Angola margin. *Geochimica et Cosmochimica Acta*, **51**: 2207–2214.
- Gustafson, L.B., and Curtis, L.W. 1983. Post-Kombolgie metasomatism at Jabiluka, NT, Australia, and its significance in the formation of high-grade uranium mineralization in Lower Proterozoic rocks. *Economic Geology*, **78**: 26–56.
- Harrison, T.M. 1990. Some observation on the interpretation of feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ results. *Chemical Geology (Isotope Geoscience section)*, **81**: 219–229.
- Harrison, T.M., and McDougall, I. 1982. The thermal significance of K-feldspar K-Ar ages inferred from ^{40}Ar – ^{39}Ar spectrum results. *Geochimica et Cosmochimica Acta*, **46**: 1811–1820.
- Hiatt, E.E., Kyser, K., Dalrymple, R.W. 1999. Sequence stratigraphy of the Proterozoic Thelon Formation, Nunavut Territory, Canada: accommodation space and systems tracts in an ancient fluvial-dominated clastic system. Geological Society of America, Abstracts with Programs, **31**: 7.
- Hoffman, P.F. 1991. Did the breakup of Laurentia turn Gondwana Land inside-out? *Science (Washington, D.C.)*, **252**: 1409–1412.
- Idnurm, M., and Giddings, J.W. 1998. Paleoproterozoic–Neoproterozoic North America–Australia link: new evidence from paleomagnetism. *Geology*, **23**: 149–152.
- Inoue, A., and Utada, M. 1983. Further investigations of a conversion series of dioctahedral mica/smectites in the Shinzan hydrothermal alteration area, northeast Japan. *Clays and Clay Minerals*, **31**: 401–412.
- Kastner, M., and Siever, R. 1979. Low temperature feldspars in sedimentary rocks. *American Journal of Science*, **279**: 435–479.
- Kawabe, I. 1978. Calculation of oxygen isotope fractionation in quartz–water system with special reference to the low temperature fractionation. *Geochimica et Cosmochimica Acta*, **42**: 613–621.
- Kotzer, T.G., and Kyser, T.K. 1995. Petrogenesis of the Proterozoic Athabasca Basin, northern Saskatchewan, Canada, and its relation to diagenesis, hydrothermal uranium mineralization and paleohydrology. *Chemical Geology*, **120**: 45–89.
- Kotzer, T.G., Kyser, T.K., and Irving, E. 1992. Paleomagnetism and evolution of fluids in the Proterozoic Athabasca Basin, northern Saskatchewan, Canada. *Canadian Journal of Earth Science*, **29**: 1474–1491.
- Kyser, T.K., and Kerrich, R.W. 1990. Geochemistry of fluids in tectonically active crustal regions. In *Fluids in tectonically active regimes of the continental crust*. Edited by B.E. Nesbitt. Mineralogical Association of Canada, Short Course Series, Vol. 18, pp. 133–230.
- Kyser, T.K., and O'Neil, J. 1984. Hydrogen isotope systematics of submarine basalts. *Geochimica et Cosmochimica Acta*, **48**: 48–53.
- Kyser, K., Hiatt, E., Renac, C., Durocher, K., Holk, G., and Deckart, K. 2000. Diagenetic fluids in Paleo- and Mesoproterozoic sedimentary basins and their implications for long protracted fluid histories. In *Fluids and basin evolution*. Edited by K. Kyser. Mineralogical Association of Canada, Short Course Series, Vol. 28, pp. 225–262.
- Land, L., and Dutton, S. 1978. Cementation of Pennsylvanian deltaic sandstone: isotopic data. *Journal of Sedimentary Petrology*, **48**: 1167–1176.
- Lawrence, S.R., and Taylor, H.P., Jr. 1972. Hydrogen and oxygen isotope systematics in weathering profiles. *Geochimica et Cosmochimica Acta*, **36**: 1377–1393.
- LeCheminant, A.N., and Heaman, L.M. 1989. Mackenzie igneous events, Canada: Middle Proterozoic hot-spot magmatism associated with ocean opening. *Earth and Planetary Science Letters*, **96**: 38–49.
- LeCheminant, A.N., Ashton, K.E., Chiarenzelli, J., Donaldson, J.A., Best, M.A., Tella, S., and Thompson, D.L. 1983. Geology of Aberdeen Lake map area, District of Keewatin: preliminary report. Current research, part A, Geological Survey of Canada, Paper 83-1A, pp. 437–448.
- Liu, K., and Epstein, S. 1984. The hydrogen isotope fractionation between kaolinite and water. *Chemical Geology, Isotope Geoscience*, **2**: 335–350.
- Marumo, K., Nagasawa, K., and Kuroda, Y. 1980. Mineralogy and hydrogen isotope geochemistry of clay minerals in the Ohnuma geothermal area, northeastern Japan. *Earth and Planetary Science Letters*, **47**: 255–262.
- McCrea, J.M. 1950. On the isotopic chemistry of carbonates and paleotemperature scale. *Journal Chemical Physics*, **18**: 849–857.
- Miller, A.R. 1995. Polymetallic unconformity-related uranium veins in lower proterozoic Amer group, Pelly Lake map area, northern Thelon Basin, Churchill Province, Northwest Territories. In *Current Research 1995-C*; Geological Survey of Canada: 151–161.
- Miller, A.R., and LeCheminant, A.N. 1981. Geology and uranium metallogeny of Proterozoic supracrustal successions, central district of Keewatin, N.W.T. with comparison to northern Saskatchewan. In *Geology of Uranium Deposits*. Edited by T.I.I. Sibbald and W. Petruk. Proceedings of the Canadian Institute of Mining (CIM) –

- Society of Economic Geologists (SEG) Uranium symposium, p. 32.
- Miller, A.R., Cumming, G.L., and Krstic, D. 1989. U–Pb, Pb–Pb, and K–Ar isotopic study and petrography of uraniferous phosphate-bearing rocks in the Thelon Formation, Dubawnt Group, Northwest Territories, Canada. *Canadian Journal of Earth Science*, **26**: 867–880.
- Page, R.W. 1988. Geochronology of early to middle Proterozoic fold belts in northern Australia: a review. *Precambrian Research*, **40/41**: 1–19.
- Pagel, M., and Jaffrezic, H. 1977. Analyses chimiques des saumures des inclusions du quartz et de la dolomie du gisement d'uranium de Rabbit lake (Canada). Aspect méthodologique et importance génétique. *Comptes Rendus des Séances de l'Académie des Sciences, Série D*, **284**: 389–414.
- Pagel, M., Michard, A., Juteau, M., and Turpin, L. 1993. Sm–Nd, Pb–Pb, and Rb–Sr systematics of the basement in the Cigar Lake area, Saskatchewan, Canada. *Canadian Journal Earth Science*, **30**: 731–742.
- Philippe, S., Lancelot, J.R., Clauer, N., and Pacquet, A. 1993. Formation of the Cigar Lake uranium deposit based on U–Pb and K–Ar isotope systematics. *Canadian Journal Earth Science*, **30**: 720–730.
- Ramaekers, P. 1990. Geology of the Athabasca Group (Helikian) in northern Saskatchewan. Saskatchewan Geological Survey (SGS) Report 195.
- Reynolds, R.C., Jr. 1985. NEWMOD a computer program for the calculation of one-dimensional diffraction patterns of mixed-layered clays. R.C. Reynolds, Jr., Hanover, N.H.
- Roedder, E. 1984. Fluid inclusions. *Mineralogical Society of America, Reviews in Mineralogy* 12, p. 644.
- Ross, G.M., Parrish, R.R., and Winston, D. 1992. Provenance and U–Pb geochronology of the Mesoproterozoic Belt Supergroup (northwestern United States): implications for age of deposition and pre-Panthalassa plate reconstructions. *Earth Planetary Science Letters*, **113**: 57–76.
- Ruzicka, V. 1996. Unconformity-associated uranium. In *Geology of Canadian mineral deposit types. Edited by O.R. Eckstrand, W.D. Sinclair, and R.I. Thorpe. Geological Survey of Canada, Geology of Canada, No. 8*, pp. 197–210.
- Savin, S.M., and Lee, M. 1988. Isotopic studies of phyllosilicates. In *Hydrous phyllosilicate (exclusive of micas). Edited by S.W. Bailey. Mineralogical Society of America, Reviews in Mineralogy* 19, 189–223.
- Sheppard, S.M.F., and Schwarcz, H.P. 1970. Fractionation of carbon and oxygen isotopes and magnesium between coexisting metamorphic calcite and dolomite. *Contributions to Mineralogy and Petrology*, **26**: 161–168.
- Taylor, H.P. 1974. The applications oxygen and hydrogen studies to problems of hydrothermal alteration and ore deposition. *Economic Geology*, **69**: 843–883.
- Wilde, A.R., Mernagh, T.P., Bloom, M.S., and Hoffmann, C.F. 1989. Fluid inclusion evidence on the origin of some Australian unconformity-related uranium deposits. *Economic Geology*, **84**: 1627–1642.
- Wilson, M.R., and Kyser, T.K. 1987. Stable isotope geochemistry of alteration associated with the Key Lake uranium deposit. *Economic Geology*, **82**: 1450–1557.
- Wright, G.M. 1955. Geological notes on central District of Keewatin, NWT. Geological Survey of Canada, Paper 55, p. 17.
- Yeh, H.W. 1980. D/H ratios and late-stage dehydration of shales during burial. *Geochimica et Cosmochimica Acta*, **44**: 341–352.
- Zhang, Y., and Frantz, J.D. 1987. Determination of the homogenization temperatures and densities of supercritical fluids in the system NaCl–KCl–CaCl₂–H₂O using synthetic fluid inclusions. *Chemical Geology*, **64**: 335–350.