

Basin evolution, diagenesis and uranium mineralization in the Paleoproterozoic Thelon Basin, Nunavut, Canada

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ABSTRACT

The Paleoproterozoic (Statherian) Thelon Basin is located in the Churchill Province of the Canadian Shield, formed following the Trans-Hudson Orogeny. Basin formation followed an interval of felsic volcanism and weathering of underlying bedrock. The diagenetic evolution of the Thelon lasted about one billion years and was punctuated by fluid movement influenced by tectonic events. Early quartz cements formed in well-sorted, quartz-rich facies during diagenetic stage 1; fluids in which these overgrowths formed had $\delta^{18}\text{O}$ values near 0‰ (Vienna Standard Mean Ocean Water). Uranium-rich apatite cement (P1) also formed during diagenetic stage 1 indicating that oxygenated, uranium-bearing pore water was present in the basin early in its diagenetic history. Syntaxial quartz cement (Q1) formed in water with $\delta^{18}\text{O}$ from -4 to -0.8 ‰ in diagenetic stage 2. Diagenetic stage 3 occurred when the Thelon Formation was at *ca.* 5 km depth, and was marked by extensive illitization, alteration of detrital grains, and uranium mineralization. Basin-wide, illite crystallized at ~ 200 °C by fluids with $\delta^{18}\text{O}$ values of 5–9‰ and δD values of -60 to -31 ‰, consistent with evolved basinal brines. Tectonism caused by the accretion of Nena at *ca.* 1600 Ma may have provided the mechanism for brine movement during deep burial. Diagenetic stage 4 is associated with fracturing and emplacement of mafic dikes at *ca.* 1300 Ma, quartz cement (Q3) in fractures and vugs, further illitization, and recrystallization of uraninite (U2). Q3 cements have fluid inclusions that suggest variable salinities, $\delta^{18}\text{O}$ values of 1.5–9‰, and δD values of -97 to -83 ‰ for stage 4 brines. K-feldspar and Mg-chlorite formed during diagenetic stage 5 at *ca.* 1000 Ma in upper stratigraphic sequences, and in the west. These phases precipitated from low-temperature, isotopically distinct fluids. Their distribution indicates that the basin hydrostratigraphy remained partitioned for > 600 Ma.

INTRODUCTION

The Paleoproterozoic (Statherian) Thelon Basin (Fig. 1) is an intracratonic sedimentary basin in the Churchill Province of the Canadian Shield. This basin, like the contemporaneous Athabasca Basin to the southwest (Fig. 1) and other continental basins of the Baraboo interval (1750–1650 Ma), contains a thick basal clastic succession that experienced burial to *ca.* 5 km and a protracted fluid history (Kyser *et al.*, 2000; Renac *et al.*, 2002). Proterozoic continental basins differed from younger Phanerozoic analogs in that they were virtually free of mud-rich facies (e.g. Gall *et al.*, 1992; Hiatt *et al.*, 2003), and did not contain terrestrial plant matter (cf. Schumm, 1968); these are key differences

that make Proterozoic basins favorable for unconformity-type uranium mineralization (Kyser & Hiatt, 2003). Silt and clay-sized sediment would have been widely eroded by eolian and fluvial processes without land plants to trap and bind them (e.g. Schumm, 1968). Absence of land plants also meant that Proterozoic continental basins were devoid of organic material capable of making pore fluid reducing. U is mobile in oxidizing pore water; Phanerozoic analogs usually contain plant-derived organic matter that acts as a diffuse reducing agent trapping U and preventing it from being concentrated. The sandstones and conglomerates that make up the basin-filling Thelon Formation have not been metamorphosed, are flat lying, and were affected by relatively low-temperature processes for more than one billion years (Renac *et al.*, 2002). These rocks provide an excellent opportunity to study the long-term effects of diagenesis, pore fluid evolution, and possible driving mechanisms for fluid movement in a Proterozoic continental sedimentary basin.

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Sandstone-dominated continental basins, common in the Paleoproterozoic, host economic metal deposits such as unconformity-type uranium deposits that formed during burial diagenesis (e.g. Hoeve & Sibbald, 1978; Wilson & Kyser, 1987; Ruzicka, 1993). The conventional approach to understanding many of these deposits is to examine ore characteristics and proximal alteration (e.g. Ruzicka, 1993). However, holistic basin analysis techniques, used in petroleum reservoir evaluation, can also be adapted to understand the large-scale fluid fluxes involved in the generation of ore deposits in Proterozoic basins (Kyser & Hiatt, 2003; Kyser, 2007).

Previous studies of the Thelon Basin have concentrated on understanding sedimentology and lithostratigraphy (Donaldson, 1965; Cecile, 1973; Jackson *et al.*, 1984; Gall *et al.*, 1992; Rainbird & Hadlari, 2000; Rainbird *et al.*, 2003), diagenetic fluid history and mineral paragenesis (Renac *et al.*, 2002), and sequence stratigraphy and hydrostratigraphy (Hiatt *et al.*, 2003; Rainbird *et al.*, 2003; Palmer *et al.*, 2004). Important questions remain regarding the evolution and fluid history of the Thelon Basin, including: (1) what drove fluid movement responsible for diagenetic events, including uranium mineralization, (2) how did these events relate to overall basin evolution, (3) what makes Proterozoic sedimentary basins, like the Thelon, amenable to mineralizing events that produced large unconformity-type uranium deposits?

Herein, we integrate mineral paragenesis, temperatures for diagenetic phases, and absolute ages determined by Renac *et al.* (2002), with a refined paragenesis, stable isotopic compositions of authigenic phases, calculated pore water isotopic compositions, and lithostratigraphic and hydrostratigraphic models to delineate the basin evolution and fluid history for the Thelon Basin. The purpose of this study was to (1) constrain important diagenetic phases within a regional sequence stratigraphic framework; (2) determine the nature and distribution of fluids associated with diagenetic minerals and (3) provide a temporal and spatial framework in which mechanisms that drove burial diagenesis and fluid flow in this Paleoproterozoic basin can be understood.

GEOLOGIC SETTING

The Paleoproterozoic (2500–1600 Ma) was marked by several major orogens associated with assemblage of Arctica (Canada, Siberia and parts of Greenland) and Atlantica (Africa and South America; e.g. Rogers, 1996). The Statherian period (1800–1600 Ma; Ogg, 2004) was a time of widespread continental sedimentary basin formation and deposition of thick clastic sedimentary rock successions interbedded with mostly felsic volcanics across North America (e.g. Holm *et al.*, 1998; Karlstrom *et al.*, 2001; Rainbird *et al.*, 2003; Medaris *et al.*, 2007). The Thelon Basin formed following the Trans-Hudson orogen (2000–1800 Ma) but before the major tectonic events associated with the growth of Arctica, which occurred near the end

of the Paleoproterozoic through the beginning of the Mesoproterozoic (1700–1500 Ma) with the accretion of Baltica, Laurentia and East Antarctica into the larger continent of Nena (Rogers, 1996). Atlantica and other continental fragments were further accreted to Nena during the Grenville event (*ca.* 1.0 Ga) to form the supercontinent Rodinia (e.g. Karlstrom *et al.*, 2001). Near the end of the Proterozoic, continental-scale tectonism occurred once again as the mega-continent Rodinia broke into several fragments (e.g. Rogers, 1996).

The Thelon Basin is a coarse clastic sedimentary rock-dominated intracratonic basin that lies within the Rae Domain of the Western Churchill Province (Fig. 1). The Thelon Formation (Barrenland Group) is the dominant basin-filling unit, has a maximum stratigraphic thickness of 2 km, and consists of quartz-dominated sandstones and conglomerates (Fig. 2; Donaldson, 1965; Gall *et al.*, 1992). The Thelon Formation was deposited on a deeply weathered paleoregolith suggesting a significant depositional hiatus (Gall, 1994). It is overlain by a thin basaltic volcanic unit, the Kuungmi Formation, and dolostones of the Lookout Point Formation (Fig. 2; Gall *et al.*, 1992; Rainbird *et al.*, 2003). The latter formations are exposed only to the west of the study area where access is restricted because outcrop exposures are located in the Thelon Game Sanctuary (Fig. 1).

The absolute age for onset of basin formation and Thelon deposition is not known. Rainbird & Davis (2007) found that primary zircons from the youngest rhyolite flows in the underlying Pitz Formation date to 1753 ± 2 Ma. Basal conglomerate and the saprolitic basement are locally cemented and crosscut by fluoro-apatite veins (Miller *et al.*, 1989). U–Pb dates of 1720 ± 6 to 1685 ± 4 Ma for this apatite (Miller *et al.*, 1989), and the age of the underlying Pitz Formation (Rainbird *et al.*, 2003) constrain the onset of Thelon Formation deposition to between 1753 and 1720 Ma.

Hiatt *et al.* (2003) divided the Thelon Formation into three stratigraphic sequences (Fig. 2). Rainbird *et al.* (2003) added an additional sequence that includes the overlying Kuungmi and Lookout Point formations. Four major lithofacies occur in varying proportions within each sequence (Fig. 2; Hiatt *et al.*, 2003). These include immature coarse sediments interpreted as alluvial fan deposits, fining-upward packages of poorly to moderately sorted sandstone and conglomerate beds capped by thin fine-grained layers that reflect high- to low-energy braided fluvial channel deposits, and well-sorted quartz arenites interpreted as upper shoreface and eolian facies (Jackson *et al.*, 1984; Hiatt *et al.*, 2003).

Sediment was derived dominantly from the east and northeast (Fraser *et al.*, 1970; Palmer *et al.*, 2004) but paleocurrents at the southern margin of the basin suggest flow from the south (Rainbird & Hadlari, 2000). Analysis of detrital clasts, including detrital zircon ages, showed that the source area changed through time. Sequence 1 (Fig. 2) is dominated by proximally derived material; sequence 2, based on ages of detrital zircons, contains sediment

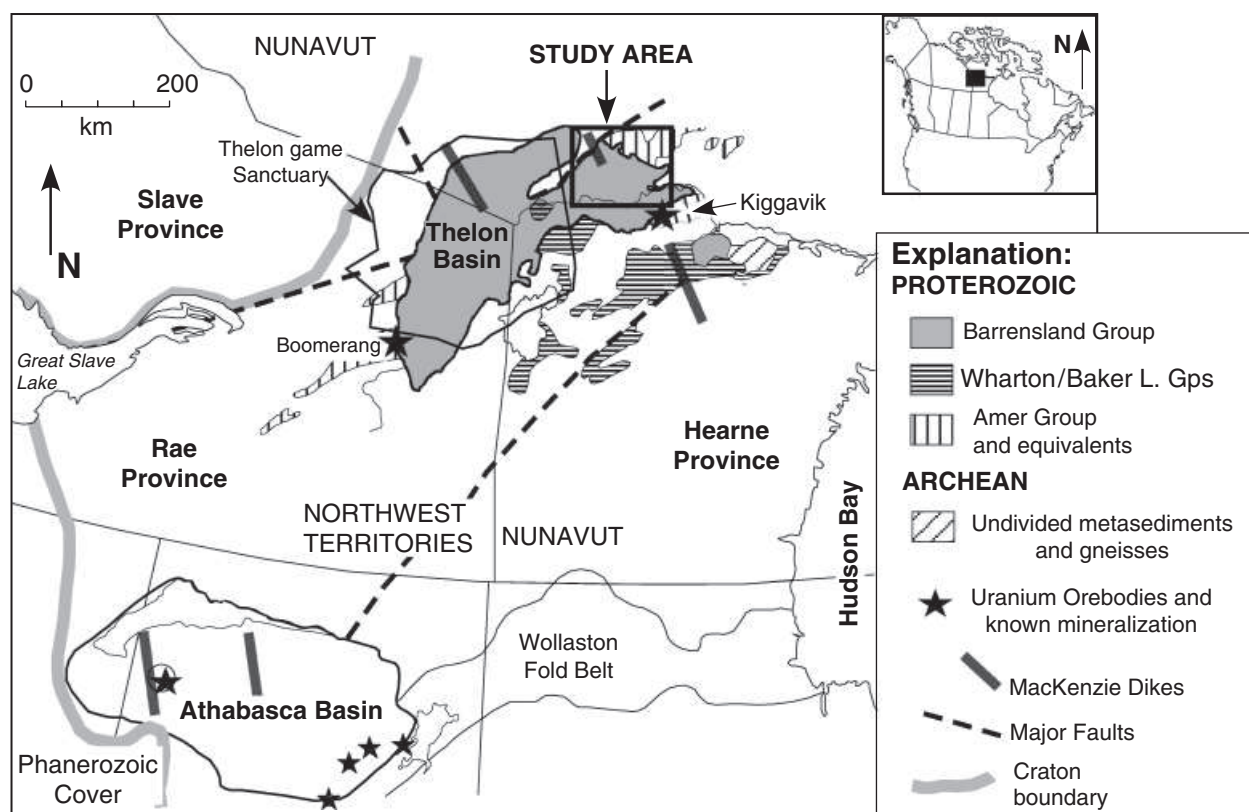


Fig. 1. Regional map showing the study area, western Canadian Shield, position of the Paleoproterozoic Thelon and Athabasca basins and their relationship to major structural features, selected Proterozoic and Archean rocks, and uranium mineralization. Modified from Gandhi (1989).

derived from as much as 2000 km away and sequence 3 again contains more proximally derived sediment (Palmer *et al.*, 2004). Thelon Formation detrital material consists primarily of quartz (95–98%) with minor feldspar (most altered to some degree by illite) and lithic fragments (schist, banded iron formation, Pitz volcanics, Amer Group metapelite, quartzite), trace amounts of detrital muscovite, biotite, oxides, zircon, apatite, tourmaline and rutile, and locally abundant, phosphatic mud intraclasts. Detrital zircons are concentrated in heavy mineral bands that often contain hematite pseudomorphs after magnetite grains (Renac *et al.*, 2002; Rainbird *et al.*, 2003).

Stratigraphic correlation indicates significant topography and synsedimentary faulting affected sedimentation in the Thelon Basin. Hiatt *et al.* (2003) documented *ca.* 25 m of paleotopographic relief on the basal unconformity in the northern part of the sub-basin during initial deposition of the Thelon sandstone, and late fault-related offset of *ca.* 350 m.

Uranium mineralization

Uranium mineralization in the Thelon Basin consists of deposits on the eastern edge of the basin and showings on the periphery (Fuchs *et al.*, 1986; Davidson & Gandhi, 1989; Fuchs & Hilger, 1989; Gandhi, 1989). Mineralization was discovered at Boomerang Lake on the southwestern basin

margin and substantial deposits at Kiggavik 5 km south of the study area (Fig. 1). Boomerang Lake, the only known mineralization hosted within the Thelon Formation (Fig. 1), contains subeconomic U–Au and is interpreted as a complex unconformity-type deposit with mineralogical, structural and geological associations comparable to high-grade deposits such as Cigar Lake and Rabbit Lake in the Athabasca Basin (Davidson & Gandhi, 1989; Gandhi, 1989).

The Kiggavik–Sissons–Schultz deposits consist of three separate basement-hosted unconformity uranium deposits near the original Kiggavik discovery 80 km west of Baker Lake, Nunavut (64°26'N and 97°40'W; Figs 1 and 3; AREVA, 2008). The original Kiggavik deposit contains 15 554 metric tons at 0.274% as U metal; and the nearby Sissons–Schultz deposits contain 22 754 metric tons U metal at 0.377%, and 13 298 metric tons at 0.126% U metal, for a total of 51 606 metric tons at an average grade of 0.26% U metal in the three known Kiggavik deposits (AREVA, 2008). The deposits lie 2–4 km beyond the present-day eastern extent of the Thelon Basin (Fig. 1) and presumably were once unconformably overlain by the Thelon Formation. The Kiggavik deposits are hosted in flat-lying quartzites of the Ketyet River Group near a *ca.* 1700 Ma Hudsonian granite pluton and crosscut by late, *ca.* 1270 Ma, diabase intrusions that postdate mineralization (Fuchs *et al.*, 1986). The uranium mineralization occurs as

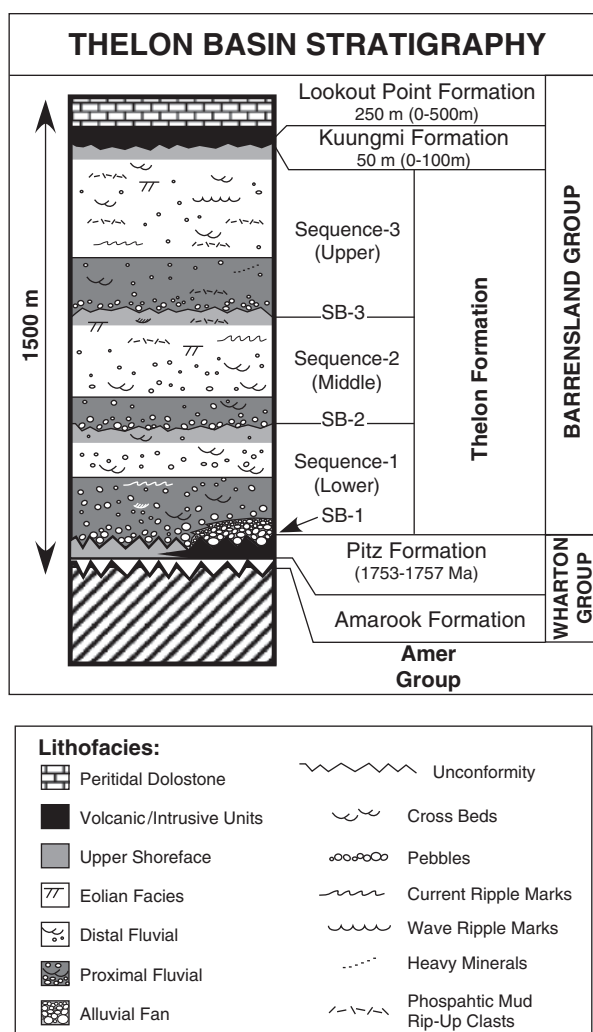


Fig. 2. Lithologic and stratigraphic relationships for the Thelon Basin based on Gall *et al.* (1992), Rainbird & Hadlari (2000), and Hiatt *et al.* (2003). SB, sequence boundary; dates for Pitz Formation are from Rainbird & Davis (2007).

plunging, elongate pods located at the intersection of two major deposit-scale faults/shear zones; one trending east northeast (065° Azimuth) dipping south southeast intersecting an east–west trending fault dipping northward (Fuchs & Hilger, 1989). Mineralization in the form of pitchblende and coffinite occurs as colloform aggregates close to fractures and finely disseminated granules associated with altered pyrite grains (Fuchs & Hilger, 1989). Pyrite associated with mineralization is largely altered to hematite and occurs with illite formed by hydrothermal processes (Fuchs *et al.*, 1986). Although there are no direct absolute ages for the uranium minerals, associated hydrothermal illite was dated using whole-rock K–Ar at 1648 and 1563 Ma in relatively unaltered samples, and at 1358 and 1073 Ma from altered rock bearing uranium mineralization, reflecting later recrystallization (Fuchs *et al.*, 1986). This suggests that uranium-bearing pore water was migrating in the basin as early as *ca.* 70 Ma after onset of basin formation.

MATERIALS AND METHODS

Samples of Thelon Formation were collected from 12 drill cores and three outcrops throughout the eastern Thelon Basin (Fig. 3). The distribution and timing of authigenic phases have been determined from petrographic study of 200 thin sections, X-ray diffraction of mineral separates, and clay mineralogy was determined using a Portable Infrared Mineral Analyzer (PIMA; Integrated Spectronics Pty. Ltd., Castle Hill, New South Wales, Australia) on 1690 samples (Appendix S1). Polished thin sections were examined to establish the volume and genetic relationships of authigenic cements, and to locate primary fluid inclusions. Thin sections were impregnated under vacuum with blue-dyed epoxy in order to estimate the amount of pore space, including microporosity.

Transmitted-light and cold-cathodoluminescence petrography (CL) were used to study diagenetic phases. CL microscopy was used to characterize quartz generations based on luminescence characteristics and completed using a Technosyn model 8200 Mark II cathodoluminescent system (Technosyn Ltd, Cambridge, UK) operating at 20-mtorr vacuum, 14–16 kV, 450–550 μ A and exposure times of 10–15 min. In addition to CL, fluid inclusion and stable isotope compositions of well-documented quartz and clay cement phases (Renac *et al.*, 2002; and this study) were used to indicate the character and distribution of fluids responsible for diagenesis in specific lithologies. Intergranular volumes were estimated from visual estimates of grains, pore space and of early-formed cements based on study of 200 thin sections.

Stable isotope compositions of quartz and clay cements were measured at Queen's University using conventional BrF_5 techniques of Clayton & Mayeda (1963) for oxygen and of Kyser & O'Neil (1984) for hydrogen and are reported in units of per mil (‰) relative to Vienna Standard Mean Ocean Water. The stable isotope compositions of clay minerals were determined from 2 to 5 μm separates obtained by ultrasonic agitation of clays from gently hand-crushed whole-rock samples. The compositions of the clay separates were determined by X-ray diffraction before stable isotope analysis. Oxygen isotope compositions of fluids that precipitated the clays were calculated using the fractionation factors of Savin & Lee (1988) for illite–water and chlorite–water. Hydrogen isotope compositions of fluids were calculated using the fractionation equation of Marumo *et al.* (1980) for chlorite–water and approximated for illite using the smectite–water fractionation equation of Yeh (1980). Dates for illite crystallization were determined using the $^{40}\text{Ar}/^{39}\text{Ar}$ laser-heating technique of Lee *et al.* (1990).

Individual quartz crystals from veins of fracture-lining quartz (Q3) were hand-picked for stable isotope analysis. The $\delta^{18}\text{O}$ of the pore water was calculated using fluid inclusion temperature data and the quartz–water fractionation equation of Kawabe (1978). Hydrogen isotope compositions of burial fluids were measured by thermally decrepitating single-generation fluid inclusions of pure fracture-lining quartz (Q3) and reflect the original δD signature of the

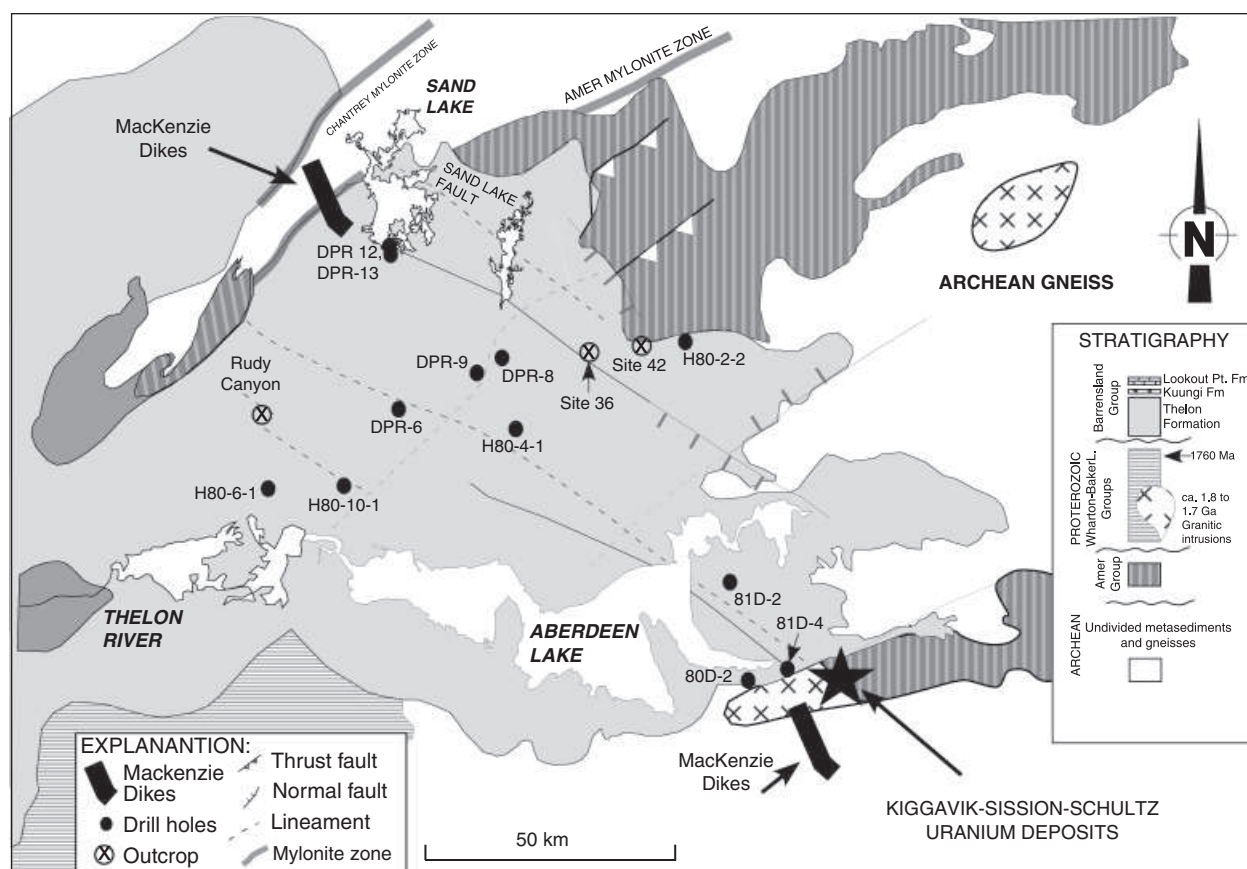


Fig. 3. Detailed geology of the eastern Thelon Basin with location of drill cores and outcrops sampled for this study. Inset shows relative stratigraphic relationships of the Thelon Formation to surrounding basement. See Fig. 1 for location.

precipitating fluids. Replicate $\delta^{18}\text{O}$ analyses were reproducible to $\pm 0.2\%$ and δD values to $\pm 3\%$. Hydrogen isotope compositions were not determined for other generations of quartz cement due to the difficulty of separating such small quantities of cement from the sample.

The temperature of quartz cementation was estimated by microthermometry of doubly polished sections containing authigenic quartz with primary fluid inclusions. Fluid inclusions were analyzed at Queen's University on a USGS-style (Fluid Inc., Broken Arrow, Oklahoma, USA) heating-freezing stage calibrated between -100 and 300°C using synthetic inclusions. Based on a geothermal gradient of $35^\circ\text{C}/\text{km}$, a 5-km maximum burial depth, which corresponds to a lithostatic pressure of 1200 bars, and a salinity of 25 wt% NaCl-equivalent, the calculated temperature correction could be as great as $\pm 55^\circ\text{C}$. However, given the likely hydrostatic pressure and salinity, the appropriate correction could be much less (Kotzer & Kyser, 1995). Because of the uncertainties in actual hydrostatic pressure and salinity, we applied no pressure corrections. Typical errors on the temperature of homogenization are within 3°C . Trace element concentrations of authigenic and diagenetic apatite were measured using a Finnigan MAT Element[®] high-resolution inductively coupled plasma mass spectrometer (ICP-MS; ThermoScientific, Bremen, Germany). Samples of authigenic apatite were drilled from within phosphatic

mud rip-up clasts using a dental drill. Samples of apatite cement were isolated by separating sandstone adjacent to the mud clasts where petrographic analyses showed apatite cement to be abundant. The rock was gently disaggregated and weighed. Both authigenic and diagenetic apatite were dissolved in 2 M nitric acid. Insoluble material was collected, dried and reweighed to determine the mass of material analyzed. Mean relative standard deviation for all elements in Table 1 is 15%.

RESULTS

Sedimentary facies in the Thelon Formation include poorly sorted polymict lithic cobble conglomerate formed in alluvial fan settings (Fig. 4a) to polymict lithic pebble conglomerates formed in proximal braided fluvial settings (Fig. 4b), and coarse-grained poorly to moderately sorted pebbly coarse-grained arenites interpreted to have formed in proximal braided fluvial systems. Moderately to well-sorted coarse-grained quartz arenite dominates the succession and is interpreted to have formed in distal braided fluvial settings that were punctuated with thin interfluvial paleosol horizons (Fig. 4c), and locally contain abundant phosphatic mud rip-up clasts (Fig. 4d) that contain high concentrations of uranium (Table 1). Moderately to well-

Table 1. Trace element concentrations of authigenic apatite intraclasts (P0) and diagenetic apatite cement (P1)

	Phase	U (p.p.m.)	Pb (p.p.m.)	Th (p.p.m.)	Sr (p.p.m.)	Mn (p.p.m.)	Cu (p.p.m.)
1-193.9-clast	P0	674	479	206	557	88	2
1-194-clast	P0	752	541	202	997	183	2
DPR6-270-clast	P0	231	110	146	187	13	1
DPR6-353.1-clast	P0	387	229	149	176	11	1
Means		511	340	176	479	73	1
1-193.9-cement	P1	1007	372	251	476	429	68
1-194-cement	P1	1068	388	969	532	1714	503
DPR6-270-cement	P1	570	113	154	292	14	3
Means		882	291	458	433	719	191

sorted medium to coarse-grained quartz arenite beds with wave ripple marks and low-angle planar wedge cross-bedding are interpreted to have formed in upper shoreface (lacustrine or marine) environments (Fig. 4e). Relatively rare, but distinctive units composed of well-sorted medium-grained quartz arenite with large-scale (individual foresets up to 4 m in length) festoon cross-bedding are interpreted to have formed in eolian dune complexes (Fig. 4f; Jackson *et al.*, 1984; Hiatt *et al.*, 2003). It is the latter three lithofacies that are well-sorted and quartz-rich, that preferentially experienced early quartz cementation, and record the first major basin-scale diagenetic event in the Thelon Basin (Hiatt *et al.*, 2007). Based on relative timing, absolute ages, and cross-cutting relationships, the authigenic minerals occur in five discrete stages during the burial history of the Thelon Basin (Fig. 5).

Diagenetic stage 1

Diagenetic stage 1 began soon after deposition and lasted until *ca.* 1700 Ma; it includes distinctive quartz cement that forms isopachous rims on grains (Q_{QC}), kaolinite (K; common above the sub-Thelon Formation regolith) and apatite cement (P1; Fig. 5; Renac *et al.*, 2002). The Q_{QC} phase is a microcrystalline quartz cement within sequence-capping distal fluvial to upper shoreface, and eolian facies of the lower and upper sequences (Figs 5 and 6a–d). This cement phase is marked by prismatic to bladed quartz crystals that average 50 μ m in long dimension formed on quartz and lithic grains, and are not syntaxial with respect to the underlying grains (Fig. 6d). The Q_{QC} phase does not contain fluid inclusions large enough for analysis; it luminesces blue under CL and clearly stands out against the dull to nonluminescent detrital grains and later quartz cements (seen as light gray rims on detrital grains in Fig. 6b). Where present, these cements occupy 30–80% of the intergranular pore volume and bridged detrital grains closing off pore throats early in their diagenetic history.

Apatite and K are locally important early-formed diagenetic phases. Apatite cement is found in close proximity to authigenic apatite intraclasts (P0; Fig. 4d). The clasts luminesce red–orange in CL, and the cement crystals are

zoned dull gray to yellow–orange in CL (Fig. 6h). The mean U concentration within the phosphate cement (P1) is 882 and 511 ppm in the clasts (P0; Table 1). K (Fig. 5) is limited to facies near sequence boundaries and near the basal unconformity in the southeastern study area (Fig. 7).

Diagenetic stage 2

Q1 overgrowths represent the earliest widespread cement phase throughout the Thelon Formation and occur in fluvial, upper shoreface and eolian facies. Q1 cement, however, is most volumetrically abundant in upper shoreface facies (Fig. 8a and b) and often overgrows earlier isopachous cements (Q_{QC}). In fluvial facies, however, Q1 overgrowths on detrital grains are <50 μ m thick, are volumetrically minor comprising <5% of intergranular pore volume, and are commonly replaced by illite (Fig. 8a and b).

Q1 overgrowths have $\delta^{18}O$ compositions that range from 11.3 to 16.6 with a mean of 13.6 ± 2.0 ($n = 10$; Table 2) and formed at *ca.* 130 °C based on fluid inclusion studies of Renac *et al.* (2002). Fluids in apparent isotopic equilibrium with Q1 had $\delta^{18}O$ values of -3.4 to -0.8 with a mean of $-2.2 \pm 1.2\%$ ($n = 7$) in coarse-grained fluvial facies, and distinctly higher values of 0.6–1.9 with a mean of $1.3 \pm 0.7\%$ ($n = 3$) in upper shoreface and eolian facies (Table 2).

Breakdown and dissolution of zircon and other heavy minerals began at this stage in coarse-grained braided fluvial facies (Fig. 8c and d). Hematite pseudomorphs after heavy mineral grains, such as magnetite are common along altered heavy mineral laminae.

Diagenetic stage 3

Illite (II) is the dominant intergranular phase in the fluvial facies (Fig. 8a and b), comprises <15% of intergranular volume in the upper shoreface facies, and is nearly absent in the eolian facies. II occurs as pore-filling fibrous crystals, or as radial cements that show little evidence of compaction. Ar^{40} – Ar^{39} ages of II range from 1650 Ma in well-cemented upper shoreface facies to 1400 Ma in distal fluvial facies (Renac *et al.*, 2002). II replaced both detrital and Q1 cement as evidenced by cusped boundaries and

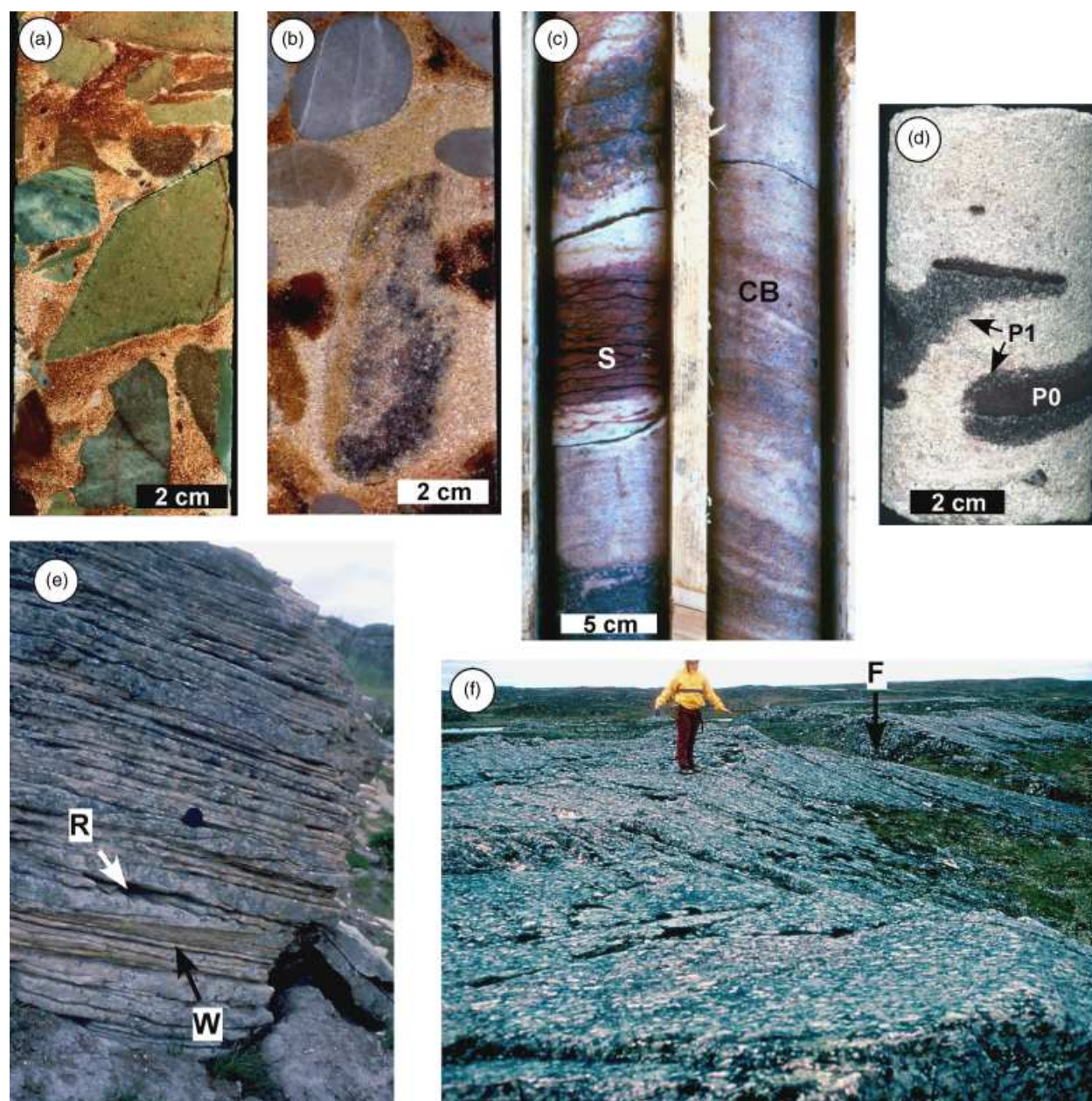


Fig. 4. Examples of lithofacies found in the Thelon Formation. (a) Slabbed drill core from core DPR-12 2 m above the basal unconformity composed of lithic pebble conglomerate from the alluvial fan facies. The matrix is coarse sand with abundant feldspar grains altered to illite; the rock was a diagenetic aquifer during peak diagenesis. (b) Sample of lithic pebble to cobble conglomerate of the proximal braided fluvial facies (diagenetic aquifer) from drill core DPR-12. (c) Unslabbed drill core showing interfluve paleosol horizon (S) and cross bedding (CB) typical of the distal braided fluvial facies, Drill core DPR-6. (d) Phosphatic mud rip-up intraclasts in coarse white quartz arenites of distal braided fluvial facies. Apatite cement (P1) forms haloes around the intraclasts and helped retain hematite; most of the hematite was leached during water–rock interaction. (e) Outcrop exposure of thinly bedded well-sorted medium-grained quartz arenite of the upper shoreface lithofacies at Rudy Canyon (RC) locality. The rock is well cemented by Q1 and is a diagenetic aquitard. Lens cap in center of photo is 3.5 cm in diameter; R, ripplemark; W, wedge cross bedding. (f) Lichen-covered outcrop of medium-grained well-sorted quartz arenites with large-scale, festoon cross bedding (F) of the eolian facies, site 42 locality (Fig. 3). The rock is extensively cemented with Q_{QC} and Q1 quartz cements and is a diagenetic aquitard.

remnant Q1 (Fig. 8a and b). This replacement is widespread suggesting illitization occurred in silica-undersaturated pore water. Il has δD values that range from -76 to -48 with a mean of $-63 \pm 8\text{‰}$, $n = 42$, and $\delta^{18}\text{O}$ values that range from 11.3 to 16.2 with a mean of $13.4 \pm 1.2\text{‰}$, $n = 33$ (Table 3).

$\delta^{18}\text{O}$ values of water in equilibrium with Il in the poorly sorted, poorly cemented fluvial facies range from 5.2 to 8.8 with a mean of $7.1 \pm 1.1\text{‰}$, $n = 20$, and those of the well-sorted quartz arenite of the upper shoreface and eolian facies range from 5.7 to 10.1 with a mean of $7.7 \pm 1.4\text{‰}$, $n = 13$ (Table 3; Fig. 9a). Water in apparent equilibrium with Il in

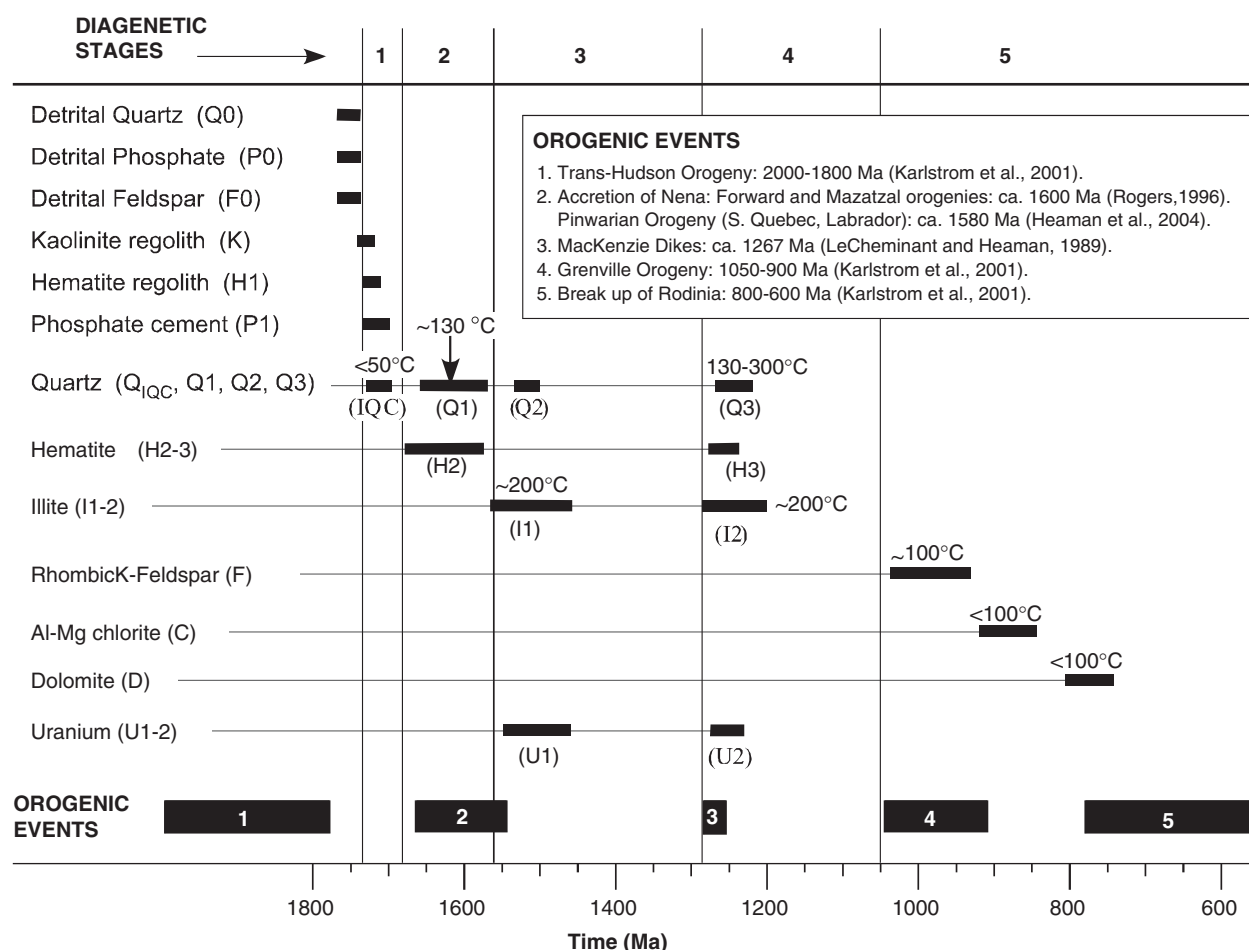


Fig. 5. Paragenetic relationships for the Thelon Basin based on Renac *et al.* (2002) and Hiatt *et al.* (2007) with major tectonic events added (numbered black boxes).

the fluvial facies had δD values that range from -60 to -31 with a mean of $-48 \pm 8\%$, $n = 29$, and those of the well-sorted, well cemented lithofacies have δD values ranging from -60 to -32 with a mean of $-46 \pm 8\%$, $n = 13$, assuming equilibrium at a temperature of 200°C [temperature determined from illite crystallinity measurements by Renac *et al.* (2002); Fig. 9b].

A minor quartz cement phase (Q2) formed locally along fractures during this stage. Q2 is distinguished from Q1 and from Q_{QC} because it overprints illite and is easily distinguished from microcrystalline cement by its larger crystal size, association with fractures and its nonluminescence under CL.

Diagenetic stage 4

Coarse-grained illite (I2) is concentrated along sequence boundaries. I2 is much less abundant than I1 and formed later, at 1300–1150 Ma (Renac *et al.*, 2002).

Q3 quartz cement is a fracture-lining to filling phase comprised of coarse drusy crystals up to 1 cm in length that have primary fluid inclusions indicating a temperature of 140°C (Table 4). Q3 cement crystals have $\delta^{18}\text{O}$ values of 16.2–18.6 with a mean of $17.7 \pm 0.8\%$, $n = 7$, and

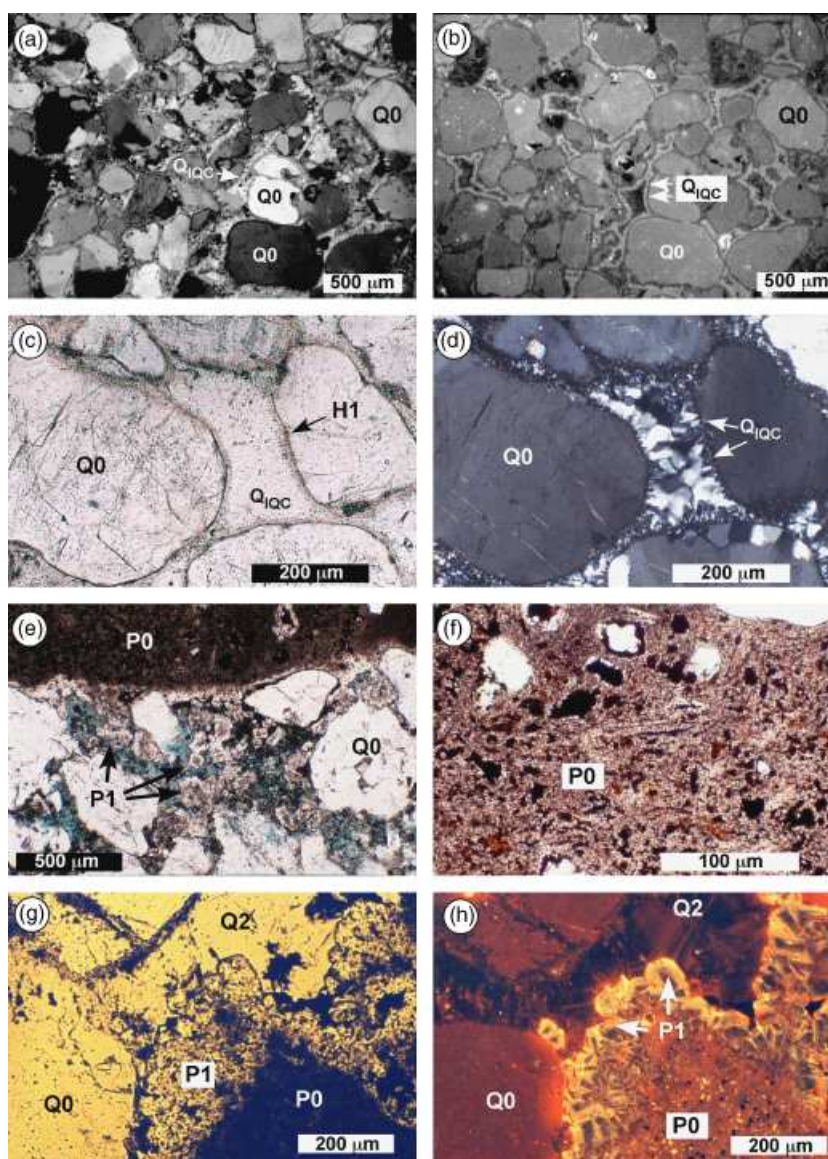
their isotopic signatures vary regionally (Table 2). $\delta^{18}\text{O}$ values of fluids that formed Q3 at 140°C range from 1.5 to 3.9 with a mean of $3.0 \pm 0.8\%$, $n = 7$ (Table 2). Fluid from which Q3 precipitated had δD values ranging from -97 to -83 with a mean of $92 \pm 8\%$, $n = 3$ (Table 2) that are distinct from basin-wide brines responsible for earlier quartz and illite (I1). Homogenization temperatures range from 102 to $>364^\circ\text{C}$, but the majority of crystals analyzed are in the range 102–247 $^\circ\text{C}$ with a mean of $144 \pm 41^\circ\text{C}$, $n = 26$. Four samples produced temperatures $>350^\circ\text{C}$ and all are associated with faults and fractures.

Diagenetic stage 5

Late diagenetic phases include K-feldspar (F; 1–5% of intergranular pore volume), Al-Mg-chlorite (C) and dolomite (D; $<1\%$ of intergranular pore volume; Fig. 5). C sometimes replaced Q1 and I1 and can occur as pseudomorphs after illite (Fig. 8e and f). In the central portion of the study area, C often accounts for $>50\%$ of intergranular pore volume (Fig. 7).

Chlorite is the dominant pore-filling clay in sequence 3 and throughout cores from the northwestern portion of the study area (Fig. 7) and formed at ca. 75°C based on its

Fig. 6. (a) Photomicrograph in plane-polarized light (PPL) showing early-formed isopachous nonsyntaxial quartz cement (Q_{IQC}) developed on detrital quartz ($Q0$) in well-sorted quartz arenite of the eolian facies, Site 42 (Fig. 3), (b) Same field of view as in (a) in CL. Q_{IQC} luminesces blue, which appears as light gray rims around $Q0$ here. (c) Photomicrograph in PPL showing higher magnification view of Q_{IQC} cement and well developed hematite rims (H1) shown in (a and b). (d) Same field of view as in C in cross-polarized light (XPL). (e) Photomicrograph in PPL showing the edge of a phosphate intraclast (P0) from sample shown in Fig. 4d. Apatite cement (P1) is inclusion-rich, zoned and forms halos around the intraclasts. (f) Photomicrograph in PPL showing closer view of internal texture in a phosphate intraclast (P0). Note the complex pattern of crystallites that could represent ghost textures of cyanobacteria filaments. Dark areas are hematite. (g) Photomicrograph in PPL showing apatite cement (P1) and quartz cement Q2. (h) Photomicrograph in CL showing zonation in P1 from dull gray cores to bright yellow luminescing rims. Q2 overlying P1 is nonluminescent and Q0 is dark red.



crystal chemistry and crystallinity (Renac *et al.*, 2002). Chlorite has δD values that range from -69 to -48 with a mean of $-57 \pm 7\%$, $n = 13$, and $\delta^{18}O$ that ranges from 10.7 to 14.0 with a mean of 12.2 ± 0.8 , $n = 17$ (Table 3). Fluid in equilibrium with chlorite had δD values that range from -41 to -20 with a mean of $-29 \pm 7\%$, $n = 13$, and $\delta^{18}O$ values from -5.3 to -2.0 with a mean of $-3.8 \pm 0.8\%$, $n = 17$ (Table 3).

Dolomite is the youngest preserved authigenic phase and only occurs in sequences 1 and 2 (Fig. 7). Poikilitic

dolomite cement (D; Fig. 5) is present most commonly within 10–20 m of sequence boundary 2 in cores DPR-6, H-80-4-1 and H80-2-2, or near boundaries with well-cemented lithologies (diagenetic aquitards; Hiatt *et al.*, 2003), where they make up $< 1\%$ of the intergranular pore volume. This phase contains fluids trapped at about $100^\circ C$ and has calculated $\delta^{18}O$ values of -6 to -4% (Renac *et al.*, 2002); it is isotopically distinct from both early basinal brines and Q3 precipitating fluids, but similar to those that formed C.

Fig. 7. Stratigraphic sections included in this study (Fig. 3) showing distribution of major diagenetic phases and their relation to lithology, sequence and hydrostratigraphy stratigraphy. Relative clay abundance normalized to 100% is shown in shaded columns to the left of stratigraphic logs. Ar/Ar ages of illite are shown in the stratigraphic position of where samples were collected. Dates with white lettering on black background represent those collected from diagenetic aquitards, and those with black lettering are from diagenetic aquifers. The 1263 Ma date from near the base of core H80-4-1 is from a sample on the unconformity. SB, sequence boundary; K, kaolinite (diagenetic stage 1); F, K-feldspar cement (diagenetic stage 4); and D, dolomite cement (diagenetic stage 5).

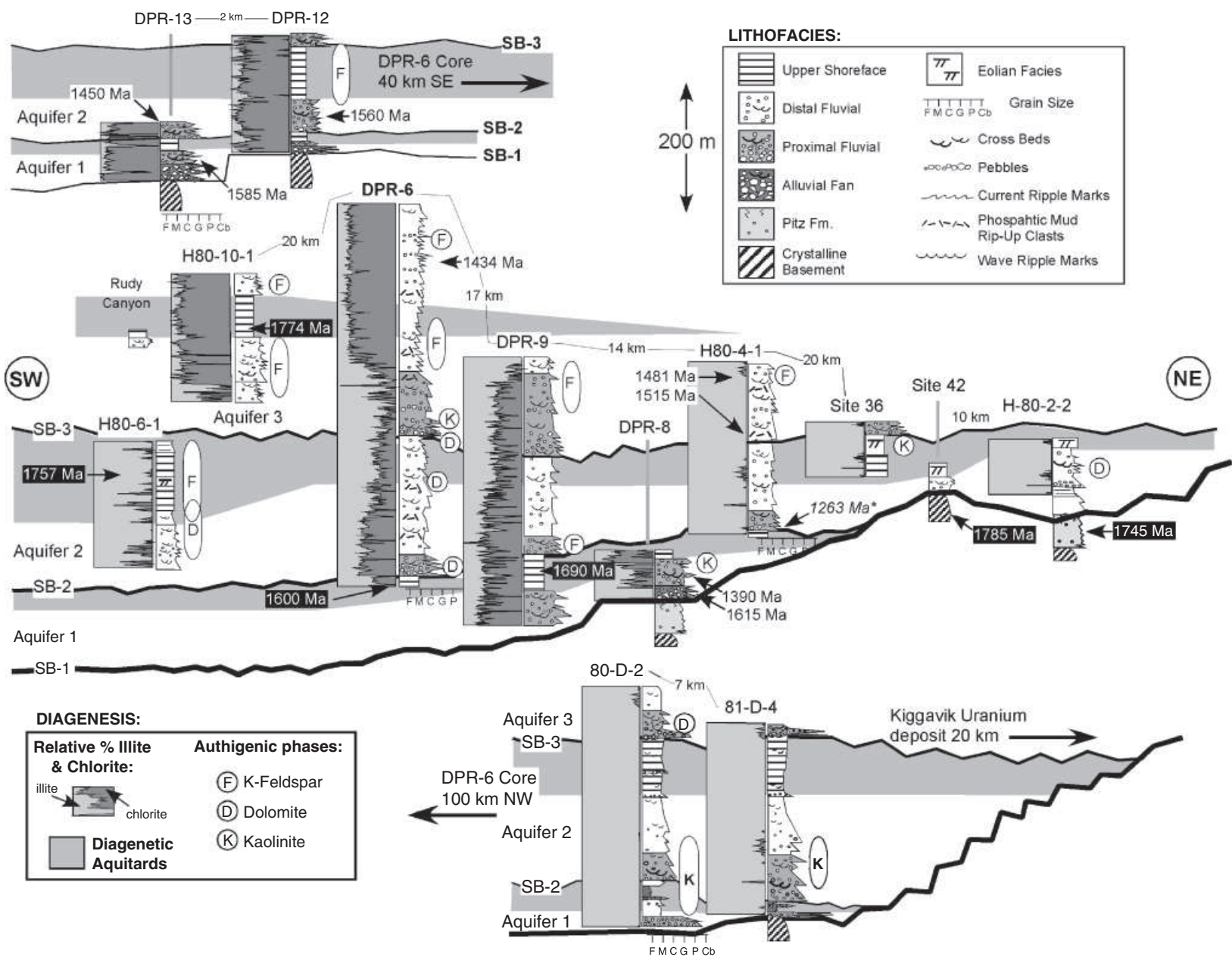


Fig. 8. (a) Photomicrograph in PPL showing Q1 overgrowths on detrital quartz grains (Q0) in the coarse-grained fluvial facies in core DRP-6. Illite (I1) fills intergranular pore space and corroded Q1 rims; some of the hematite dust rims (H1) on Q0 are preserved. (b) Same field of view as (a), but in XPL showing cusped ragged border between I1 and Q1. (c) Photomicrograph in XPL showing a badly altered detrital zircon typical of those found in diagenetic aquifers, middle sequence, core DPR-8. (d) Photomicrograph in reflected light showing a partially dissolved detrital zircon in the proximal braided fluvial facies of the upper sequence in core DPR-6, 175 m. (e) Photomicrograph in PPL from a quartz wacke from the lower sequence in core DPR-6 in which intense pressure solution of Q0 has occurred. A detrital zircon is relatively well preserved and is surrounded by chlorite (C) and I1. (f) Same view as in (e), but in XPL. Note that I1 replaced the detrital quartz grain in the middle of the field of view. Chlorite later replaced the illite and probably also corroded the Q0 grain.

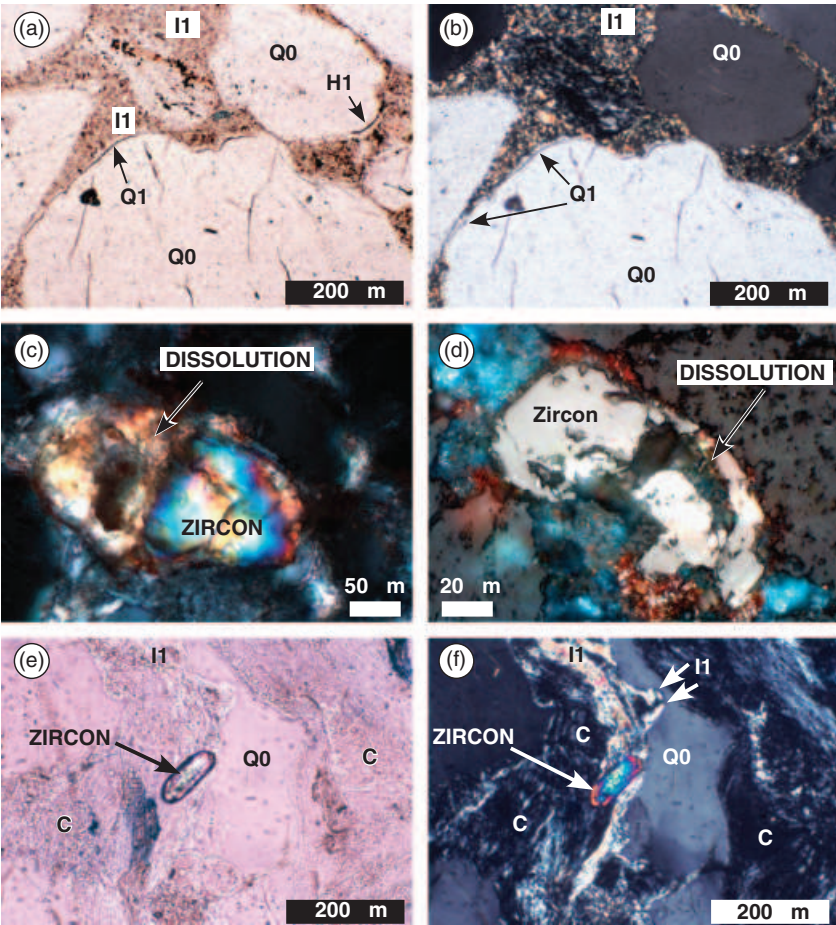


Table 2. $\delta^{18}\text{O}$ values of quartz cement overgrowths (Q1), and $\delta^{18}\text{O}$ and δD_w of fracture-lining quartz (Q3) in the Thelon Formation						
Sample	Depth	Phase	Unit	$\delta^{18}\text{O}$ qtz	δD_w	$\delta^{18}\text{O}_w$
H80-10-1	31	Q1	3, FL	13.5	ND	-1.2
	39	Q1	3, FL	11.3	ND	-3.4
	135	Q1	3, AT	15.3	ND	0.6
	163	Q1	3, FL	11.3	ND	-3.4
	205	Q1	3, FL	12.6	ND	-2.1
81-D-4	96	Q1	2, AT	16.6	ND	1.9
H80-6-1	62	Q1	2, AT	16.1	ND	1.4
	172	Q1	2, FL	11.3	ND	-3.4
H80-4-1	150	Q1	2, FL	13.8	ND	-0.9
	210	Q1	2, FL	13.9	ND	-0.8
Rudy Canyon (outcrop)	OC	Q3	3, AT	18.6	-83	3.9
	OC	Q3	3, AT	18.5	ND	3.8
	OC	Q3	3, AT	18.1	ND	3.4
DPR-12	192	Q3	2, FL	17.4	-97	2.7
	233	Q3	1, AT	16.2	ND	1.5
DPR-9	440	Q3	1, AT	17.5	-95	2.8
	456	Q3	1, AT	17.7	ND	3.0

Measured δD_w values and calculated $\delta^{18}\text{O}_w$ of fluids in apparent equilibrium with these phases based on temperatures of 130 °C, as derived from fluid inclusion data of Renac *et al.* (2002) and this study, are also shown. The fractionation equation of Kawabe (1978) was used to calculate $\delta^{18}\text{O}_w$. Stratigraphic unit definitions: 1, 2 and 3 represent the lower sequence, middle sequence and upper sequence of Hiatt *et al.* (2003), respectively. FL, coarse-grained braided fluvial facies that were diagenetic aquifers; AT, diagenetic aquitard formed in well sorted upper shoreface to eolian facies. OC, outcrop sample; ND, no data.

Table 3. $\delta^{18}\text{O}$ and δD values of authigenic clays (II and C), and $\delta^{18}\text{O}_w$ and δD_w values of fluids in apparent equilibrium with these phases based on temperatures of 200 and 75 °C, respectively; temperatures were determined by Renac *et al.* (2002)

Sample	Depth	Unit	Min	$\delta D_{\min}$	$\delta^{18}\text{O}_{\min}$	δD_w	$\delta^{18}\text{O}_w$
Sequence 3 (upper)							
H80-10-1	31*	FL	C	– 51	12	– 23	– 4.0
	34*	FL	C	– 50	11.8	– 22	– 4.2
	39*	FL	C	– 69	10.7	– 41	– 5.3
	93*	AT	C	– 48	11.6	– 20	– 4.4
	114*	AT	C	– 52	11.3	– 24	– 4.7
	163*	FL	C	– 54	11.9	– 26	– 4.1
	205*	FL	C	– 55	12	– 27	– 4.0
H80-6-1	22*	FL	II	– 72	14.2	– 56	8.1
H80-4-1	70*	FL	II	– 61	12.8	– 45	6.7
	90*	FL	II	– 70	13.2	– 54	7.1
	110*	FL	II	– 63	11.3	– 47	5.2
DPR-6	25	FL	C	– 68	ND	– 40	ND
	96	FL	C	– 64	12.8	– 36	– 3.2
	124	FL	C	ND	12.1	ND	– 3.9
	124	FL	C	ND	12.6	ND	– 3.4
	124	FL	C	ND	12.0	ND	– 4.0
	193	FL	C	ND	12.1	ND	– 3.9
	193	FL	C	ND	12.7	ND	– 3.3
	244	AT	C	– 60	13.2	– 32	– 2.8
	251	AT	II	ND	11.8	ND	5.7
	251	AT	II	ND	11.9	ND	5.8
	270	AT	C	ND	13.4	ND	– 2.6
	270	AT	C	ND	14.0	ND	– 2.0
	309	AT	II	– 54	13.1	– 37	7.0
	318	AT	C	– 59	11.3	– 31	– 4.7
	322	FL	II	– 65	13.6	– 48	7.5
	351	FL	II	– 67	14.6	– 51	8.5
Rudy Canyon	OC	AT	II	– 73	13.3	– 57	7.2
Sequence 2 (middle)							
81-D-4	43.7*	AT	II	– 48	16.2	– 32	10.1
	96*	AT	II	– 64	15.4	– 48	9.3
	225*	FL	II	– 60	14.6	– 44	8.5
	260*	FL	II	– 54	13.2	– 38	7.1
	266*	FL	II	– 50	14.4	– 34	8.3
DPR-8	32	FL	II	– 69	ND	– 53	ND
	32	FL	II	– 68	ND	– 52	ND
DPR-9	217	AT	II	– 69	ND	– 53	ND
	237	AT	II	– 62	ND	– 46	ND
	359	AT	II	– 61	ND	– 45	ND
DPR-6	521	FL	II	– 50	12.8	– 33	6.7
	521	FL	II	– 54	ND	– 38	ND
	563	FL	II	– 55	12.1	– 39	6.0
	600	FL	II	– 48	12.0	– 31	5.9
	616	FL	II	– 53	ND	– 36	ND
H80-6-1	42*	AT	II	– 62	13.8	– 46	7.7
	62*	AT	II	– 66	15	– 50	8.9
	94.7*	AT	II	ND	15	ND	8.9
	131*	AT	II	– 63	13.2	– 47	7.1
	172*	FL	II	– 66	14.9	– 50	8.8
	202*	FL	II	– 65	14.3	– 49	8.2
	207*	FL	II	– 58	13.5	– 42	7.4
	226*	FL	II	– 64	13.2	– 48	7.1
Sequence 2 (middle)							
DPR 13	69	FL	II	– 73	ND	– 57	ND
H-80-2-2	26*	AT	II	– 76	13.2	– 60	7.1

Table 3. (Continued)

Sample	Depth	Unit	Min	$\delta D_{\min}$	$\delta^{18}\text{O}_{\min}$	δD_w	$\delta^{18}\text{O}_w$
	72.5*	FL	II	– 76	13	– 60	6.9
	94*	FL	II	– 71	12.4	– 55	6.3
	136.4*	FL	II	– 70	11.6	– 54	5.5
	136.4*	FL	II	– 71	11.7	– 55	5.6
Sequence 1 (lower)							
DPR-6	633	AT	II	– 53	13.0	– 37	6.9
DPR-8	74	FL	II	– 72	ND	– 56	ND
	92	FL	II	– 74	ND	– 58	ND
DPR-9	415	FL	II	– 76	ND	– 60	ND
DPR-12	205	FL	C	– 59	ND	– 31	ND
	233	FL	C	– 55	ND	– 27	ND
DPR 13	108	FL	II	– 62	ND	– 46	ND
81-D-4	322.8*	AT	II	– 57	14.8	– 41	8.7

The fractionation equations of Savin & Lee (1988) were used to calculate $\delta^{18}\text{O}_w$ and δD_w .

*Data from Renac *et al.* (2002).

Stratigraphic units: FL, diagenetic aquifers; AT, diagenetic aquitards; OC, outcrop sample; ND, no data.

DISCUSSION

Paragenetic relationships and calculated diagenetic temperatures show a progression from early near-surface diagenesis progressing to deep burial diagenesis followed by uplift. Plotting burial temperatures with their corresponding estimated depths and absolute ages (from Renac *et al.*, 2002), where available, produced a burial history curve for the Thelon Basin (Fig. 10). There are five distinct stages of diagenesis, most of which correspond in time to major tectonic events that affected the craton, and had the potential to create a hydraulic head in the basin during its long burial history. Basin formation followed the Trans-Hudson Orogeny from 2000 to 1800 Ma, but smaller-scale tectonic disturbances due to rejuvenation in the Trans-Hudson region occurred as late as 1750–1735 Ma (Henderson *et al.*, 1990) approximately coeval with basin formation. The Thelon Basin was buried to about 5 km by about 1620 Ma probably as a result of tectonism associated with accretion of Nena, and probably remained deeply buried at about 200 °C for approximately 500 million years (Fig. 10).

The sedimentology and stratigraphic architecture of the basin-filling Thelon Formation established the foundation for the basin's diagenetic evolution by creating a well-defined hydrostratigraphy (Hiatt *et al.*, 2003) that, based on data presented here, persisted and affected the stratigraphic and regional extent of later burial diagenetic events through diagenetic stage 5 over 800 million years later. The stages of diagenetic evolution are summarized below.

Diagenetic stage 1

Diagenetic stage 1 is marked by near-surface diagenesis that includes quartz cementation, phosphate mobilization

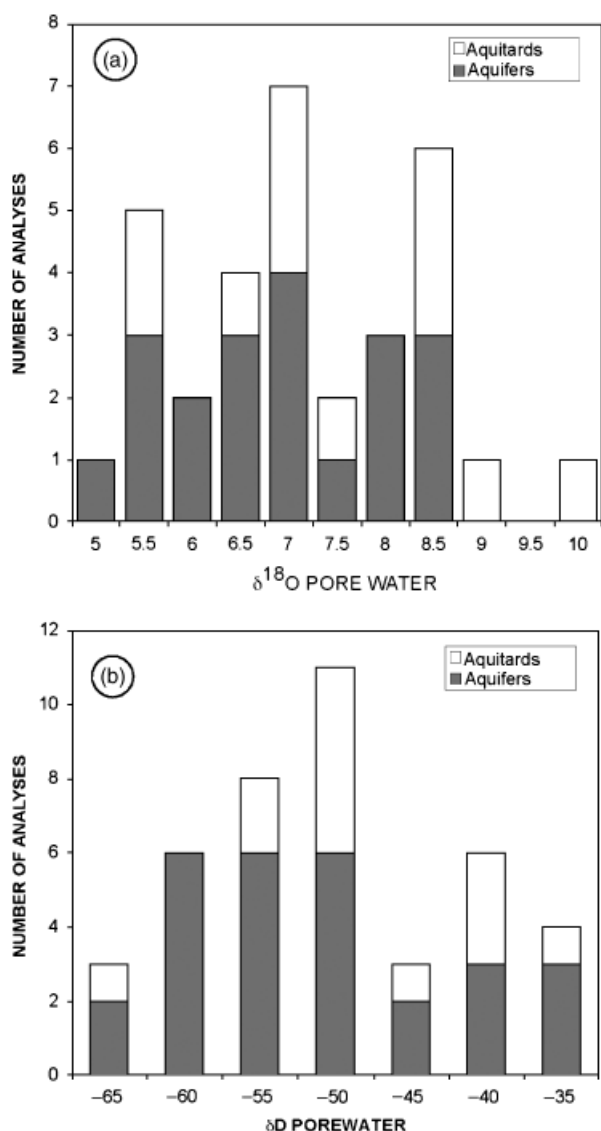


Fig. 9. (a) Distribution of $\delta^{18}\text{O}_w$ results for pore water in equilibrium with illite II in diagenetic aquitard and diagenetic aquifers units. (b) Distribution of δD_w results for pore water in equilibrium with II by diagenetic aquitard and aquifer units.

from detrital intraclasts that precipitated as apatite cement, and K (Fig. 5). Within well-sorted sandstones of the Thelon Formation, early fluids formed quartz cements in the near-surface environment (Hiatt *et al.*, 2007) and continued cementation created regional diagenetic aquitards. This occurred in two stages, initially at low temperatures as isopachous morphologies (Q_{QC}), characteristic of near-surface phreatic-zone diagenesis (Hiatt *et al.*, 2007), and later as syntaxial overgrowths during early compaction of the sediment (Renac *et al.*, 2002; Hiatt *et al.*, 2003, 2007). Q_{QC} has $\delta^{18}\text{O}$ values of 25–33‰ and represents early, low-temperature ($< 50^\circ\text{C}$), near-surface cementation that lithified sand below the water table before mechanical compaction (Hiatt *et al.*, 2007). The timing and extent of authigenic quartz precipitation at different stages during the evolution of the Thelon Basin had a significant effect

on the later paleohydrology and diagenesis of the basin-filling clastic rocks.

Authigenic apatite intraclasts (P0; Fig. 4d) that occur in the Thelon Formation represent phosphatic muds deposited in lakes or in quiet water marginal marine settings that were later ripped up and deposited as fluvial down-cutting occurred. The ultimate source of the phosphorus was likely from weathering of granitic and volcanic sediments (Gall & Donaldson, 2006). These clasts are most abundant in sequences 1 and 2 (Fig. 7) and in the center of the study area, range from 3 mm to 4.5 cm in length, and are concentrated on bedding planes in distal braided fluvial facies. The clasts luminesce dark red–gray in CL (Fig. 6h). Apatite cement crystals (P1) were recognized as hydroxyapatite by Renac *et al.* (2002) and rim the intraclasts (Fig. 4d) suggesting they are the result of precipitation of phosphate mobilized from the clasts. The cements are subhedral to euhedral and are zoned dull gray to yellow–orange in CL (Fig. 6h) suggesting fluctuating trace element concentrations during growth. Uranium concentrations (Table 1) are distinctly higher in the cements (mean of 882 ppm) relative to the clasts (mean of 511 ppm) indicating that U was present in oxidizing diagenetic pore waters early in the diagenetic history of the Thelon Formation. Based on U/Pb dating Miller *et al.* (1989) found that P1 cement formed at 1685 and 1647 Ma, and suggested that it resulted from remobilization of earlier formed vein-filling phosphate minerals, but petrographic relationships clearly show an association with phosphatic rip-up clasts (Fig. 6e–h).

K (Figs 5 and 7) is concentrated near sequence boundaries and is most abundant near the basal unconformity in the southeastern portion of the study area (Fig. 7). K may be the product of feldspar weathering and diagenesis on the exposed surfaces, which was later reworked and deposited in topographic lows and likely underwent recrystallization during burial.

Diagenetic stage 2

Diagenetic stage 2 includes syntaxial quartz overgrowths (QI) that are present throughout the Thelon Formation (Fig. 8a and b) and are abundant in upper shoreface and eolian facies. This stage represents the onset of burial diagenesis, which may have been caused by tectonism associated with the accretion of the supercontinent Nena (Figs 5 and 10). Fluids that precipitated QI had $\delta^{18}\text{O}$ values of 0.6–1.9‰ within well-cemented upper shoreface and eolian lithofacies but distinctly lower values from -4 to -0.8 ‰ in poorly sorted proximal fluvial facies (Table 3). The lower $\delta^{18}\text{O}$ values of fluids that precipitated QI cements within proximal fluvial facies relative to the more permeable distal fluvial, upper shoreface and eolian facies requires either that the fluids in the two facies differed or the temperatures of fluids that precipitated the cement in the proximal fluvial facies were higher than those of the well-sorted distal fluvial to upper shoreface facies. If the difference were due to temperature, cements in the diagenetic aquitards would have formed at 20–30 °C lower than

Table 4. Description, size, temperature of first melting (T_{fm}), ice melting (T_{ice}), homogenization temperatures (T_{h}) and salinities in wt% equivalent NaCl from fluid inclusion analysis of fracture-lining quartz (Q3), and euhedral quartz (Q2) of the Thelon Formation and quartz veins in the underlying basement (Pitz Formation)

Sample	Phase	Description	Size (μm)	T_{fm}	T_{ice}	T_{h}	wt% eq NaCl
Thelon Formation							
DPR-12-192	Q3	(l+v+s)	14	ND	ND	101.7	ND
	Q3	(l+v+s)	37	ND	− 6.8	121.1	10
	Q3	(l+v+s)	38	− 63.9	− 28.4	123.4	> 30
	Q3	(l+v+s)	32	ND	ND	124.6	ND
	Q3	(l+v+s)	35	ND	ND	126.2	ND
	Q3	(l+v+s)	17	ND	ND	126.4	ND
	Q3	(l+v+s)	35	ND	ND	126.5	ND
	Q3	(l+v+s)	39	− 70.7	− 25.9	127.8	> 30
	Q3	(l+v+s)	12	ND	ND	129.9	ND
	Q3	(l+v+s)	18	ND	− 16.1	133.7	20
	Q3	(l+v+s)	15	− 47	ND	136.2	ND
	Q3	(l+v+s)	38	ND	− 5.6	136.6	9
RC-2	Q3	(l+v)	20	− 71.8	ND	118.2	ND
	Q3	(l+v)	34	− 68	ND	119.7	ND
	Q3	(l+v)	18	ND	ND	120.0	ND
	Q3	(l+v+s)	12	ND	ND	125.4	ND
	Q3	(l+v+s)	10	ND	ND	134.1	ND
	Q3	(l+v+s)	37	− 68.8	ND	137.7	ND
RC-1	Q3	(l+v+s)	14	ND	ND	146.0	ND
	Q3	(l+v)	10	ND	ND	117.2	ND
	Q3	(l+v)	8	ND	ND	129.7	ND
	Q3	(l+v)	8	ND	ND	133.0	ND
	Q3	(l+v)	10	ND	ND	225.7	ND
	Q3	(l+v)	10	ND	ND	232.3	ND
	Q3	(l+v)	9	ND	− 27.6	236.3	> 30
	Q3	(l+v)	10	ND	− 10	247.3	12
	Q3	(l+v)	10	ND	ND	> 350	ND
	Q3	(l+v)	7	ND	ND	> 350	ND
	Q3	(l+v)	10	ND	− 15	> 350	19
	Q3	(l+v)	10	ND	− 31.3	> 364	> 30
Basement							
Pitz Fm. s.s.	Q3	(l+v+s)	20	− 41.4	− 22.3	102.5	23
	Q3	(l+v)	10	ND	− 17.3	209.0	20
	Q3	(l+v)	12	ND	− 1.1	280.0	2
	Q3	(l+v)	15	− 5.7	− 4.7	300.0	7
	Q3	(l+v)	15	ND	− 21.3	300.0	23

RC, Rudy Canyon outcrop; s.s., sandstone; l, v, s refer to the phases present in fluid inclusions, and are liquid, vapor and solid, respectively.

cements in the diagenetic aquifers. This is consistent with a hydrostratigraphic model where Q1 developed first in well-sorted lithofacies (depositional aquifers; Hiatt *et al.*, 2003) beginning during shallow burial, whereas Q1 in the diagenetic aquifers (proximal fluvial facies) formed at greater burial depths.

During initial diagenesis, the development of quartz cement in more poorly sorted facies was inhibited by abundant clays from detrital sources, mechanical infiltration of clay by groundwater, and the breakdown of feldspar in relatively immature sands. During burial these poorly sorted units retained at least some hydraulic conductivity and became diagenetic aquifers (Hiatt *et al.*, 2003). Regardless of whether Q1 formed at different temperatures or from different fluids in well-sorted facies compared with the

poorly sorted ones, the difference in $\delta^{18}\text{O}$ values reflects early stratification and isolation of fluids in distinct stratigraphic units within the basin.

Diagenetic stage 3

Diagenetic stage 3 represents peak diagenesis and was coeval coincided with the accretion of the mega-continent Nena (Baltica, North America and East Antarctica) at ca. 1600 Ma (Figs 5 and 10; Rogers, 1996). This stage is marked by widespread illitization (II), primary uranium mineralization (U1) and dissolution of Q0 and Q1. Tectonism affected large areas of the craton as reflected by the Forward and Mazatzal orogenies to the south and west, and the Pinvarian Orogeny of southern Quebec and Labrador (Cook

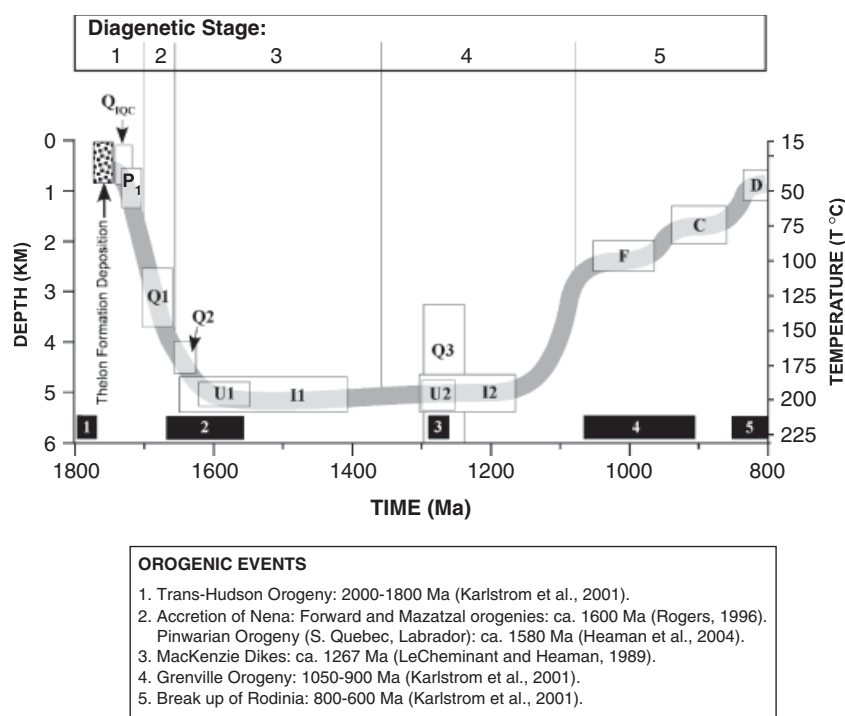


Fig. 10. Burial history curve for the Thelon Formation showing major diagenetic phases and major tectonic events. Temperature–depth scale based on temperatures determined by fluid inclusion analyses and assuming a surface temperature of 15 °C and a 35 °C/km geothermal gradient.

& MacLean, 1995; Karlstrom & Bowring, 1988; Heaman *et al.*, 2004). These far-field tectonic events affected large areas of the craton and probably played a role in causing burial of the Thelon Basin to approximately 5 km, and likely caused basin-wide fluid movement by creating a hydraulic head that, while probably small, caused movement of chemically active (*ca.* 200 °C) pore fluid through the sandstone.

Illite (II) is the most widespread and abundant diagenetic phase in the Thelon Basin (Fig. 7). II is most abundant in fluvial facies of sequences 1 and 2 and is well preserved in the southeastern drill cores (Fig. 3), and it replaces Q1 especially in diagenetic aquifers (Fig. 8a and b). II ages range from 1650 Ma in early-cemented diagenetic aquitards to 1450 Ma in diagenetic aquifers proximal to unconformities and formed at about 200 °C (Fig. 10; Renac *et al.*, 2002). Using a surface temperature of 15 °C and geothermal gradient of 35 °C/km, II would have formed at a depth of 5.3 km. The $\delta^{18}\text{O}$ values of water in apparent

equilibrium with II range from 5.2 to 10.1‰ (Table 3) and vary between stratigraphic units. The well-cemented diagenetic aquitards of sequence 2 have higher values relative to those of clay-rich, poorly sorted diagenetic aquifers (Fig. 9b). The δD values of fluids that precipitated illite (II) are variable throughout the Thelon Formation ranging from –60 to –31‰ and also vary between hydrologic units, and are slightly lower in the aquitards of sequences 1 and 2 (Fig. 9b; Table 3). This is consistent with an evolved brine and suggests that the hydrologic system remained stratified through diagenetic stage 3.

Pore water $\delta^{18}\text{O}$ values associated with the precipitation of II are the highest (> 5‰) in the southeastern drill core (81-D-4; Table 3). These higher $\delta^{18}\text{O}$ values coincide with an increase in δD values, and suggest isotopically distinct fluids were present in the southeastern portion of the basin. This area is close to known uranium mineralization at Kiggavik (Fig. 3), and elevated $\delta^{18}\text{O}$ values in II of this area may be related to ore-forming fluids.

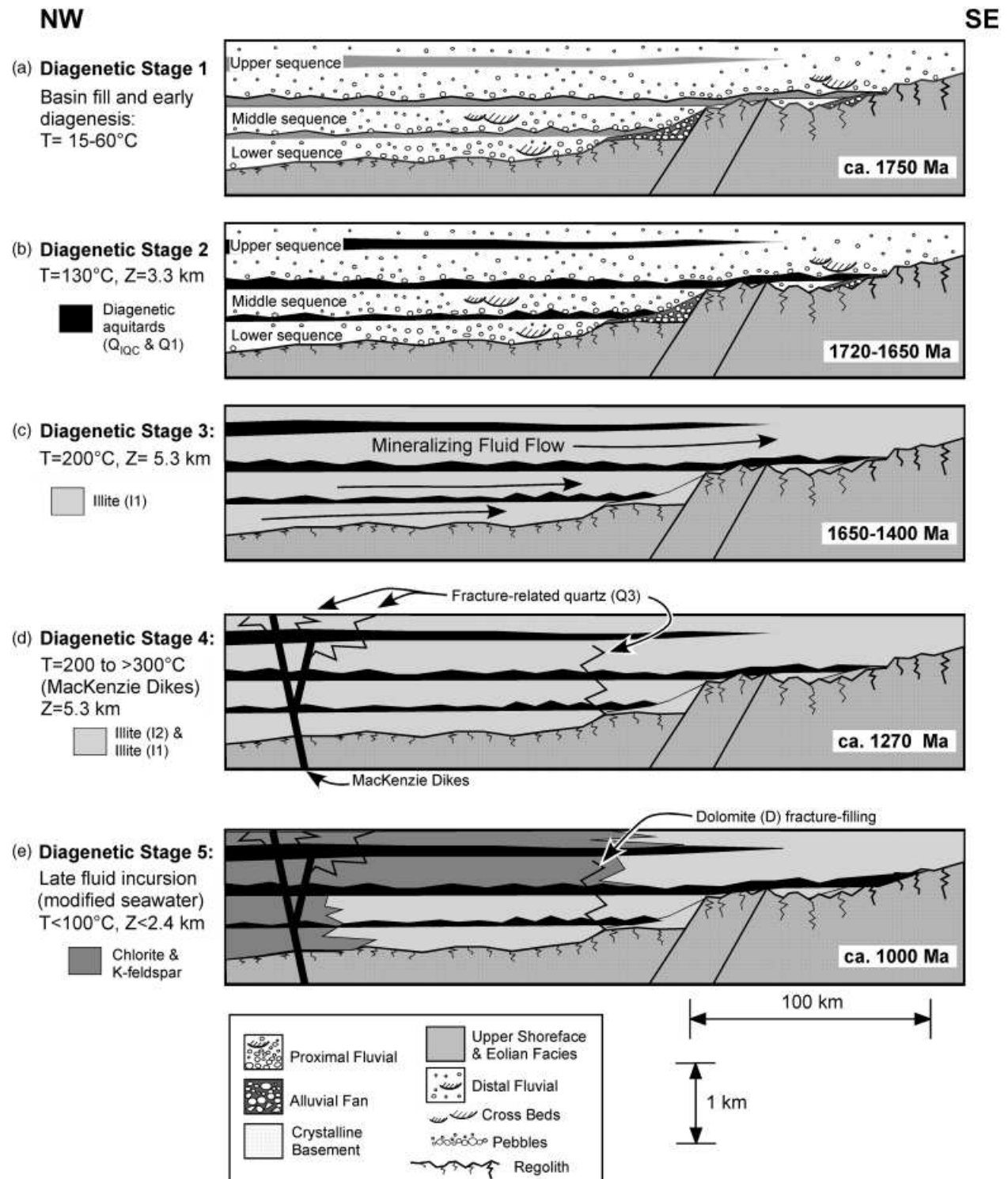
Fig. 11. Basin evolution model for the eastern Thelon Basin indicating the distribution of important authigenic diagenetic phases and basin paleo-hydrostratigraphy. (a) Diagenetic Stage 1 represents deposition to shallow burial (< 3 km). Q_{QC} cement selectively formed in well-sorted, quartz-rich upper shoreface and eolian facies. (b) Diagenetic Stage 2. Aquitards develop fully when pressure solution and Q1 cement precipitated and blocked remaining pores in upper shoreface and eolian facies. Most of Q1 formed at 130 °C and 3.3 km depth (Z). (c) Diagenetic Stage 3. Peak diagenesis occurred at *ca.* 200 °C and 5.3 km depth and is characterized by widespread precipitation of II and dissolution of quartz (Q0 and Q1) in diagenetic aquifers. The basin has entered the metal leaching window. Temperature increase caused O_2 -bearing pore water to become corrosive and transport metals to potential sites of mineralization along the basal unconformity. (d) Diagenetic Stage 4. The basin remained at about 5 km burial; incursion of hot pore water occurred along fractures associated with the MacKenzie Dikes. At this point the basin has remained at about 200 °C for > 300 Ma. Illite (I2) formed in diagenetic aquifers and remobilization of primary U1 resulted in U2 overprinting. Late stage quartz cement (Q3) lined and filled fractures. (e) Diagenetic Stage 5. Seawater-derived pore water penetrated along lithofacies boundaries producing chlorite (C) and rhombic K-feldspar (F) that partially to completely replaced existing illite in the upper and middle sequences dominantly in the northwest. Dolomite cement (D) filled sub-vertical fractures in the east.

Diagenetic stage 4

Diagenetic stage 4 is marked by uranium remobilization (U2), illitization (I2) preferentially along unconformities and in coarse-grained braided fluvial channel deposits, and locally, by coarsely crystalline quartz cement (Q3) that fills sub-vertical fractures and veins, and lines vugs. Renac *et al.* (2002) determined the age of I2 as 1200–1300 Ma and, like the earlier-formed I1, found that it formed in pore

water at *ca.* 200 °C, suggesting about 5 km burial depth. The formation of the younger I2 may represent movement of K-rich fluids due to tectonism and thermal perturbation associated with emplacement of the MacKenzie Dikes at 1267 ± 2 Ma (Figs 1, 5 and 10; LeCheminant & Heaman, 1989).

Quartz cement-filled veins (Q3) occur sporadically across the study area, but are more common to the northwest (Rudy Canyon; Fig. 3). These veins crosscut all



Thelon sequences and crosscut stylolites, Q1 and Q2 cements and illite (I1, I2) associated with peak burial diagenesis. Q3 cements are mostly found in well-cemented diagenetic aquitard facies, and there is a strong spatial correlation between Q3 and faults (Fig. 3). These veins yield a wide range of values in the fluid inclusion (Table 4) and isotopic data (Table 2). Fluid inclusions from the majority of Q3 cements indicate homogenization temperatures ranging from 100 to 140 °C, with a mean of 130 °C (Table 4). These temperatures are generally much lower than those measured for quartz veins in basement lithologies (in excess of 280 °C; Table 4). An abundance of primary, three-phase aqueous inclusions and low final ice melting temperatures from fracture-lining quartz (Q3) suggest these basinal fluids were highly saline (> 30 wt% NaCl-equivalent; Table 4). Stable isotope compositions of primary fluid inclusions in Q3 have lower δD values relative to fluids responsible for quartz (Q1) and illite (I1; Table 2).

Fluid inclusions with homogenization temperatures of ca. 130 °C coexist with those with temperatures of > 300 °C (RC locality; Table 4). The presence of high temperature fluids can be attributed to movement of fluids along faults driven by intrusions of the MacKenzie Dikes, which are pervasive throughout the region (LeCheminant & Heaman, 1989). The nearest known MacKenzie Dike exposure cuts the eastern Thelon Basin approximately 40 km to the northwest (Tella *et al.*, 1984) of Rudy Canyon near Sand Lake (Fig. 3), but it is likely that there are many more occurrences of these dikes that are not exposed at the surface. Dikes are most likely responsible for the elevated temperatures of fluids associated with precipitation of at least some of the Q3 vein-filling cements at this locality, and may indicate sporadic movement of hot pore water in the fractures.

Diagenetic stage 4 records the structural disturbance associated with emplacement of the MacKenzie Dikes which affected the region (Fig. 1). Faulting postdated diagenetic stage 3 (peak diagenesis) and caused the early-formed basin hydrostratigraphy to be breached. Q3 cement-filled fractures, while volumetrically minor on a basin scale, may have led to further compartmentalization of the hydrostratigraphy that is reflected in spatial distribution of diagenetic stage 5 phases.

Diagenetic stage 5

Diagenetic stage 5 is represented by paragenetically late K-feldspar cement (F), chlorite (C), and dolomite cement (D). Rhombic, euhedral K-feldspar cement has ^{40}Ar – ^{39}Ar ages of 1154–900 Ma (Renac *et al.*, 2002), and petrographic observations indicate they formed before chlorite (Renac *et al.*, 2002). Temperature determinations based on its crystal structure indicate formation at 100 °C or less (Renac *et al.*, 2002). The K-feldspar postdates any known U-mineralization and was likely associated with fluid movement driven by far-field effects from the Grenville Orogeny (Fig. 10).

K-feldspar (F) is restricted to sequences 3 and 2 (Figs 2 and 7) suggesting that late fluid movement was limited to the upper two sequences (Fig. 7) in the northwest part of the study area and that earlier aquitard development, in part, restricted these late basinal fluid incursions. The limited replacement of earlier illite (I1 and I2) by chlorite (C; Fig. 8e and f) in the middle and lower sequence, especially in the east and southeastern part of the study area (Fig. 7), indicates a decrease in hydraulic conductivity from upper to lower strata at this time. Thus, late stage fluids circulated more freely in the upper sequence, but did not penetrate lower sequence strata significantly, and probably did not infiltrate areas to the southeast (Fig. 7) proximal to known uranium mineralization at Kiggavik (Fig. 3).

Chlorite (C) is abundant in the upper sequence and accounts for 80–100% of the total clay content. Chlorite-dominated (C) clay groundmass is found throughout the drill core from the northwest portion of the study area (Fig. 3) and in sequences 2 and 3 of the central basin (Figs 3 and 7), but is not found in significant amounts below the diagenetic aquitard of sequence 2 or in southeast drill core (Fig. 7). Isotopic compositions of fluids that precipitated chlorite (C) have distinctly higher δD values but lower $\delta^{18}\text{O}$ values than early fluids responsible for illite (I1 and I2). Chlorite (C), and K-feldspar (F) were formed by a chemically, isotopically and temporally distinct fluid at approximately 75 °C (Renac *et al.*, 2002).

Dolomite (D) occurs predominantly in the southern part of the basin in sequences 1 and 2 near boundaries with aquitards and often is found associated with late-stage Q2 and Q3 cements (Fig. 7); overall, it is rare except where associated with Q3 cements. Petrographically, dolomite (D) is seen replacing K-feldspar (F), and calculated fluid $\delta^{18}\text{O}$ values of –6‰ and low temperatures of formation (~100 °C; Renac *et al.*, 2002) support the interpretation that D formed from a late-stage meteoric water-rich fluid that was present in the southern portion of the study area (Renac *et al.*, 2002). The more regionally comprehensive data presented here suggests that these late-stage meteoric-dominated waters did not penetrate far to the north where marine-derived pore water dominated.

The Proterozoic continental basin and unconformity-type uranium mineralization association

One of the most fundamental questions regarding the origin of unconformity-type uranium mineralization is why they are almost exclusively associated with Proterozoic continental sedimentary basins. Specifically, U is soluble in oxygenated pore water over a range of pHs and temperatures (Shock *et al.*, 1997). Uranium is mobile in pore water with an Eh above ± 0.2 and immobile under reducing conditions (e.g. Garrels & Christ, 1965; Romberger, 1985; Shock *et al.*, 1997). Oxygenated groundwater with up to 8 mg L^{–1} dissolved oxygen has been noted in modern continental sedimentary basins of the western United States to depths as great as 1 km (e.g. Winograd & Robertson,

1982). Terrestrial plant matter, however, is common in most Phanerozoic (more specifically post-Silurian) continental sedimentary basins. Decomposition of land plant-derived organic matter consumes dissolved oxygen and causes widespread diffuse reduction of U carried in oxidizing groundwater. The result is that pore water loses its ability to mobilize U and, where the water is oxidizing, leads to precipitation of U at local reduction fronts (e.g. Sanford, 1992; Spirakis, 1996), producing at best low-grade U mineralization typical of roll front and tabular stratiform sandstone-hosted U deposits (e.g. Rubin, 1970). Because of their lack of organic matter or other reduced phases such as pyrite that would have consumed dissolved oxygen, Proterozoic continental basins hosted pore water that would have had the potential to be widely oxygenated, and they would have had the potential to maintain O₂ levels at depth over space and time.

U mineralization requires U-bearing oxygenated basinal pore water interaction with reducing fluids or lithologies (Wilson & Kyser, 1987; Fayek & Kyser, 1997). Such reducing agents could have included organic matter-bearing marine sedimentary rocks, metamorphic rocks or mafic igneous rocks (e.g. Kyser *et al.*, 2000; Stefansson & Arnorsson, 2002; Kyser, 2007). These laterally extensive reducing lithologies represent effective U traps, however, interaction between U-bearing basinal brines and such lithologies results in effective trapping over a large area resulting in a diffuse relatively low concentration of U. Basinal brines that carried U require interaction with reducing fluids over laterally restricted areas and focusing along permeable zones to allow interaction with reducing basement fluids or lithologies (Fayek & Kyser, 1997); therefore favorable hydrostratigraphic relationships are critical.

Detrital grains contained in basin-filling sedimentary successions are the source of U in unconformity-related U mineralization (e.g. Hoeve & Sibbald, 1978; Wallis *et al.*, 1983; Wilson & Kyser, 1987; Fayek & Kyser, 1997; Kyser, 2007). In order to mobilize sufficient mass of U, oxidizing fluids must come in contact with a significant mass of U-bearing phases. The U-bearing phases in clastic sediments are monazite and zircon with typical U concentrations of 100s–1000s ppm, but rutile and apatite with concentrations of 10–100 ppm could contribute (Duddy & Kelly, 1999). Heavy minerals such as zircon, rutile and tourmaline account for approximately 0.04% of the total Thelon Formation and occur in all three sequences in upper shoreface and fluvial facies, but are absent from eolian facies (Palmer *et al.*, 2004).

There are significant differences in zircon source, degree of preservation, and abundance within the Thelon Formation. There is clear evidence of zircon alteration in diagenetic aquifer units (Fig. 8c and d), whereas zircon grains, where present, are much less altered in diagenetic aquitard facies, which probably accounts for their relatively high whole-rock U concentrations (Palmer *et al.*, 2004). Petrographic analysis showed that zircons were abundant in sequence 3, but they are highly altered (Fig. 8c and d; Palmer *et al.*, 2004). In se-

quence 2, preservation is variable but alteration is associated with abundance of authigenic illite (Fig. 8f). Sequence 1 generally contains relatively well-preserved zircons (Fig. 8e). This suggests that the upper sequence underwent intense water-rock interaction and that the sequences were hydraulically partitioned with little cross-sequence fluid movement when the alteration occurred.

The Thelon Formation has whole-rock U values of 0–5 ppm U (Palmer *et al.*, 2004). To date, a total of 51.6 million kilograms at 0.26% U are estimated to exist in the Kiggavik deposits (AREVA, 2008), and petrographic analysis indicates approximately 60% dissolution of heavy minerals in the diagenetic aquifers. Assuming 5 ppm for initial U content, mobilization of 60%, and diagenetic aquifer thickness of 20 m covering an area of 47 000 km², we estimate the uranium contribution by detrital grains within the basal aquifer of the Thelon Basin to be approximately 19 billion kilograms. The ratio of estimated U reserves to potential U mobilized in brines is about 0.27%. Potential errors in this calculation include over-estimates of the original U content, remobilization and dissemination of concentrated U by late fluid flow, or the presence of undiscovered deposits. Nonetheless, a significant amount of U was mobilized by brines in the basin and <1% is in currently known mineralization.

Basin evolution model for the Thelon Basin

Based on analysis of diagenetic phases in regional, sequence stratigraphic and temporal frameworks, and cross-cutting relationships, we constructed a basin evolution model for the Thelon Basin (Fig. 11). The depositional and stratigraphic architecture initially established the paleohydraulic characteristics of the basin and was a major factor determining the diagenetic pathways in the various sedimentary facies. Diagenesis is key to controlling paleohydrostratigraphy in Proterozoic continental sedimentary basins due to the lack of muddy facies (Hiatt & Kyser, 2000).

Pore water would have circulated freely in unconsolidated well-sorted quartz-rich sediments during the initial stages of basin-filling. Syn-sedimentary faulting and paleotopography would have complicated early basin hydrology, but large-scale fluid flux coupled with silica-rich pore water resulting from weathering of feldspar and volcanoclastic material (Hiatt *et al.*, 2007) in the near surface environment at this stage promoted early quartz cementation before significant mechanical compaction (Fig. 11a). Early quartz cementation was concentrated in well-sorted, quartz-rich facies creating diagenetic aquitards in the upper portions of the sequences (Fig. 11b; Hiatt *et al.*, 2007). Clay minerals are rare in aquitard units. ⁴⁰Ar–³⁹Ar analyses, however, of Illite in these units yield ages that range from 1600 to 1774 with a mean of 1705 Ma (Fig. 7) suggesting that these units were effectively closed to diagenetic fluids by 1705 Ma – before uranium mineralization occurred (ca. 1550 Ma; Fig. 5; cf. Harrison, 1990).

Compositionally and texturally immature braided fluvial facies did not experience significant quartz cementation, but are instead marked by minor syntaxial overgrowth cement that made up < 5% of the intergranular volume and formed at depths of 3–5 km ($\geq 200^\circ\text{C}$). These cements were variably replaced by illite and chlorite during burial. ^{40}Ar – ^{39}Ar analyses of illite in diagenetic aquifers yield dates that range from 1263 to 1615 Ma with a mean of 1477 Ma (Fig. 7) suggesting that these units remained open to fluid flow for at least 240 Ma after deposition of the Thelon Formation. This further suggests that diagenetic fluids were present to facilitate illite recrystallization in the diagenetic aquifers through the period of uranium mineralization and well beyond, especially along unconformities. Holk *et al.* (2003) found that radiogenic Pb produced by decay of U occurs near the basal unconformity in the eastern portion of the study area. They attributed radiogenic Pb in sequences 2 and 3 to decay of U in uranium-rich phosphate minerals (P0 and P1).

During diagenetic stages 2 and 3 (Fig. 11b and c) the rocks entered conditions in which metals could be leached and transported by warm (*ca.* 200°C) saline brines. These stages represent a period wherein the sedimentary basin entered a ‘*thermal leaching window*’ (Kyser, 2007). Rock–water interaction and mineralization require pore water flow. The question of drive or hydraulic head can only be addressed indirectly, since flow at this point in basin evolution occurs relatively late in basin history. Several major tectonic events impacted the North American craton between 1800 and 600 Ma (e.g. Rogers, 1996; Karlstrom *et al.*, 2001; Holm *et al.*, 2005). Based on correlation of diagenetic events and major regional tectonic events, it appears that these events may have caused movement of pore water in the Thelon Basin.

Late diagenetic events of stage 5 include fracture-controlled flow and incursion of pore water from the northwest that, based on comparisons of isotopic and fluid inclusion data, are consistent with evolved seawater. The overlying Lookout Point dolomite, the uppermost sedimentary unit of the Barrenland Group (Fig. 2) in the Thelon Basin, is located in the northwest portion of the basin within the Thelon Game Sanctuary (Donaldson, 1965), and attests to this basin-culminating marine stage. Another possible later source of seawater is the Neoproterozoic marine deposits of the Shaler Group that occur north of the Amundsen Basin (Ross and Kerans, 1989) and equivalent marine strata outcrop in the northern Mackenzie Mountains (Cook & Maclean, 1995). It is possible that this major marine incursion at *ca.* 1000 Ma introduced seawater to the northern Western Churchill Province, affecting permeable strata of the northern Thelon Basin.

CONCLUSIONS

- (1) The Paleoproterozoic Thelon Basin is a coarse clastic sediment-filled continental basin with a fluid history

that lasted for approximately one billion years. Basin formation and diagenesis beyond early burial were driven by large-scale far-field tectonic events that may have caused fluid movement when the Thelon Formation was buried about 5 km.

- (2) Sedimentologic and stratigraphic relationships in the Thelon established the hydrostratigraphic and diagenetic framework within the basin – a framework that influenced regional diagenetic relationships throughout the basin’s fluid history. Ar/Ar dates of illite within diagenetic aquifers and aquitards show that aquitards were closed to diagenetic fluid by *ca.* 1700 Ma, whereas aquifers remained open to diagenetic fluids until *ca.* 1480 Ma. During deep burial diagenesis, a critical time for uranium mineralization in the basin, the diagenetic aquitards focused fluids in diagenetic aquifers.
- (3) Uranium-rich apatite cement is abundant in the lower two sequences of the Thelon Formation, suggesting uranium-rich pore fluids were present early in the diagenetic history. Apatite-rich mud clasts formed in paleo-lakes provided phosphate that was precipitated around the clasts and beyond.
- (4) Diagenetic stage 3 represents a period of intense diagenesis associated with the accretion of Nena that likely caused fluid movement. It is impossible to estimate the amount of hydraulic head and direction of flow, but what is clear from the alteration of detrital minerals, basin-wide recrystallization of illite and other diagenetic evidence is that extensive water–rock interaction occurred at *ca.* 200°C over time scales of 10^7 years.
- (5) Isotopic compositions of illite from the southeastern drill cores near the Kiggavik uranium mineralization show significantly higher $\delta^{18}\text{O}$ and δD values than laterally equivalent units toward the basin center. This suggests either higher temperatures were present along the eastern edge of the study area, or interaction of basinal brines with a distinct basement fluid that may have been critical in the formation of the uranium mineralization and base metals deposits found there.
- (6) Regional and stratigraphic distribution of diagenetic phases indicate that burial fluids moved along stratigraphically controlled pathways. Fluid pathways during late diagenesis were limited in part by regional diagenetic aquitards, indicating that early basin compartmentalization had a lasting effect on fluid circulation and diagenesis throughout the history of the basin.

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Supporting Information

Additional Supporting Information may be found in the online version of this article:

Appendix S1. Clay composition as % of total volume of clays, S/N = signal to noise.

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